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Geological Survey of Newfoundland and Labrador

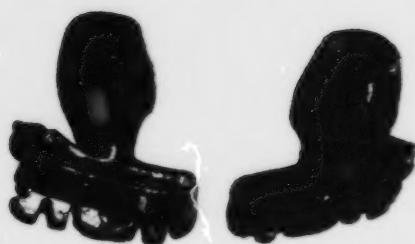
GEOLOGY AND ECONOMIC POTENTIAL OF THE PIPESTONE POND AREA (12A/1 NE; 12A/8 E), CENTRAL NEWFOUNDLAND



H. Scott Swindell

1977

Cover



(From the collection of the Newfoundland Museum)

"pipestone—..... stone, carved by the Indians into tobacco pipes."

*—Glossary of Geology, Third Edition,
American Geological Institute, Alexandria, Va.*

In Newfoundland, "pipestone" generally refers to serpentine, from which the Beothuck and Micmac Indians may have carved pipe bowls for smoking tobacco and/or willow bark. Pipestone Pond was probably named for its widespread shoreline outcrops of serpentine, part of the Pipestone Pond Ophiolite Complex.

The cover is a serpentine pipe bowl, approximately 2½ cm high. It was recovered by James P. Howley who described it thus:

".....a beautifully made pipe of greenish serpentine. The bowl is octagonal shaped outside but perfectly circular inside. There is some doubt as to whether this can be really attributed to the Beothucks, especially as they are said not to have smoked. Again, it is so fresh and unweathered, it looks as though it was made quite recently. The party who gave it to me received it from a Micmac Indian, who picked it up near Pipestone Pond in the interior and pronounced it to be of Red Indian manufacture."

—J.P. Howley, 1915

*The Beothucks or Red Indians.
Coles Publishing Co., Toronto,
page 339.*



GEOLOGICAL SURVEY BRANCH
DEPARTMENT OF MINES
GOVERNMENT OF NEWFOUNDLAND AND LABRADOR

**GEOLOGY AND ECONOMIC POTENTIAL OF THE PIPESTONE POND
AREA (12A/1 NE; 12A/8 E), CENTRAL NEWFOUNDLAND**

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Report 88-2

St. John's, Newfoundland
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ABSTRACT

The Pipestone Pond area (for the purposes of this report) comprises approximately the eastern half of the Great Burnt Lake (I2A/8) and the northeastern quadrant of the Cold Spring Pond (I2A/1) map areas. The geology of the area can be considered conveniently in terms of three fault-bounded geological domains underlain by rocks of contrasting tectonostratigraphic and/or structural and metamorphic histories. From west to east these are: 1) a western igneous-metamorphic domain underlain mainly by deformed, megacrystic granitoid rocks, but including lesser quantities of metasedimentary and metavolcanic rocks containing upper greenschist- to lower amphibolite-facies metamorphic assemblages, 2) a central domain of oceanic sedimentary, volcanic and plutonic rocks that includes an ophiolitic sequence (the Pipestone Pond Complex) and mafic to felsic volcanic and volcanoclastic sequences assigned to the Cold Spring Pond Formation and the Salmon River Dam and North Steady Pond formations of the Baie d'Espoir Group, and 3) an eastern metasedimentary domain underlain by the Spruce Brook Formation, comprising interbedded quartzite and pelite that increase progressively in metamorphic grade from greenschist in the west to upper amphibolite in the east. All stratified rocks in the area are Early to Middle Ordovician in age. Granitoid rocks in the western igneous-metamorphic domain are probably Silurian to Devonian.

The earliest structural events in the Pipestone Pond area juxtaposed the three domains along originally flat, eastward-directed thrust faults. Later reverse faults caused further tectonic interleaving of the structurally juxtaposed rocks. All rocks in the area contain principal penetrative foliations related to upright, isoclinal folds, which appear to postdate the earliest phase of thrusting. Fabrics related to later deformational events are locally developed throughout the area, but cannot be correlated on a regional scale and do not appear to be related to a single structural event.

The mafic intrusive rocks of the Pipestone Pond Complex are geochemically heterogeneous and locally show evidence of generation in a supra-subduction zone environment. Pillow lavas at the top of the ophiolite are MORB-like and not obviously part of the plutonic fractionation trend. Plagiogranites in this complex are apparently unrelated to the gabbros through any single fractionation mechanism, and are interpreted to have formed through hydrous partial melting of mafic arc crust.

A single, laterally extensive basalt unit in the Cold Spring Pond Formation, the North Salmon Dam Basalt, is interpreted to have been generated in a within-plate setting, probably in a back-arc basin. Other mafic volcanic rocks in the Cold Spring Pond and North Steady Pond formations are chemically similar to those in the Pipestone Pond Complex, and are interpreted to be fault-bounded slivers and not depositionally related to the host rocks.

Felsic volcanic rocks in the Cold Spring Pond and North Steady Pond formations are not petrogenetically related to either the Pipestone Pond Complex or the North Salmon Dam Basalt. Their geochemistry suggests an origin through melting of sialic crust, but nothing further can be postulated concerning their tectonic setting.

Geochemical and sedimentological data suggest that the central volcanic-metasedimentary-plutonic domain records the development of an Early Ordovician marginal basin. Early stages of island-arc rifting and marginal-basin formation are recorded in the Pipestone Pond Complex and the Baie d'Espoir Group. A more mature stage of basin evolution is suggested by volcanic and sedimentary facies in the Cold Spring Pond Formation. Sedimentary rocks in the western igneous-metamorphic domain are probably correlative with similar but lower grade rock types in the central domain, but detailed correlations are not possible. The Spruce Brook Formation probably represents deposition near a continental margin to the east.

Two volcanogenic massive sulphide deposits, comprising mainly pyrrhotite-pyrite-chalcopyrite, occur within the Salmon River Dam Basalt. The larger of these, the Great Burnt Lake deposit, consists of an overturned, west-facing, bedded, massive sulphide body above a footwall alteration zone. The smaller South Pond deposit comprises a sulphide-rich, alteration zone for which no exhalative equivalent has been identified. Favourable environments for ophiolitic, massive sulphide deposits are probably present in the basalts at the top of the Pipestone Pond Complex, and for volcanogenic Pb-Zn-Ag deposits in felsic volcanic environments of the Cold Spring Pond and North Steady Pond formations and the western igneous-metamorphic domain. Volcanic environments similar to gold-bearing environments in southwestern Newfoundland are present in the North Steady Pond Formation and in the western igneous-metamorphic domain. Ultramafic rocks of the Pipestone Pond Complex contain showings of chromite, asbestos and magnesite, and good opportunities exist for further discoveries of these commodities.

CHAPTER 1

INTRODUCTION

LOCATION AND ACCESS

The Pipestone Pond area, located near the geographic centre of the island of Newfoundland, includes the eastern half of the Great Burnt Lake (12A/8) map area and the northeastern quadrant of the Cold Spring Pond (12A/1) map area (Figure 1-1) (see also geological map 88-51 in back pocket with geographical names). It is bounded to the east by longitude $56^{\circ}00'$, to the north by latitude $48^{\circ}30'$, to the south by Round Pond, Godaleich Pond and the West Salmon River and to the west by an irregular, northeast-trending boundary that passes along the eastern shores of Cold Spring Pond and Great Burnt Lake.

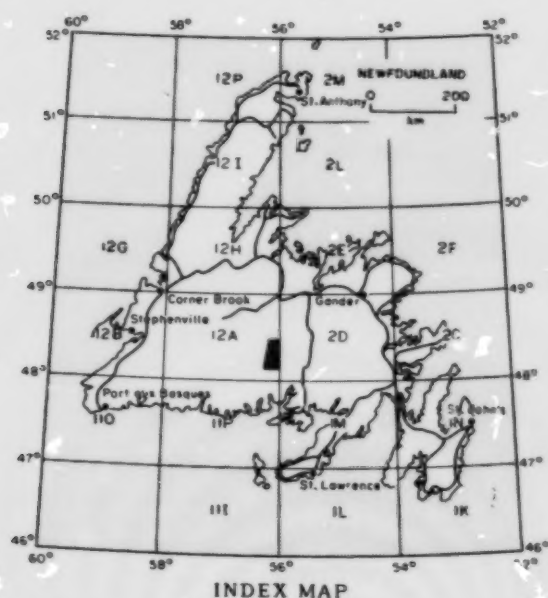


Figure 1-1: Location of the Pipestone Pond area.

There are no permanent settlements in the area. Road access to the southwestern part of the area during mapping was afforded by all-weather, gravel roads constructed to service the Upper Salmon Hydro Development. These roads connect with public routes in St. Alban's, Bay d'Espoir, approximately 50 km to the southeast, where there is a gravel airstrip, facilities for coastal shipping and daily public transport to Grand Falls. The area north and east of the North Salmon River is accessible by float plane or by helicopter. Areas adjacent to Great Burnt Lake are accessible by small boat launched from the road at the North Salmon Dam. The area north of Godaleich and the western part of Round Pond was accessible during mapping on a daily basis by canoe from the Hydro roads.

PREVIOUS WORK

The earliest systematic geological investigations in the Pipestone Pond area are attributed to Alexander Murray who ascended the Salmon River to Round Pond and then traversed the North Salmon River to Pipestone Pond and Great Burnt Lake during 1870. The geographical and geological observations of this remarkable gentleman are still definitive for the area. He recognized the principal geological units in the area, commented upon their distribution, structure and metamorphism, and noted the similarity between ultramafic sequences near Pipestone Pond and corresponding sequences in Quebec (Murray and Howley, 1881). He noted the presence of chromite, asbestos and signs of copper in the ultramafic rocks; these observations constitute the first report of concentrations of economic minerals in the area.

In 1888, J.P. Howley ascended the West Salmon River to Pipestone Pond and thence to Great Burnt Lake as part of a traverse from Bay d'Espoir to the Exploits River (Howley, 1917). His brief descriptions of the geology on the way continued Murray's previous observations. Howley (1917), was greatly impressed by the economic potential of the magnesian rocks near Pipestone Pond, which he considered to be potential sources of chromite and magnetite iron, and felt that the serpentinite and magnesian carbonate could be used as ornamental stone.

Early exploration subsequent to Murray's visit centered on the chromite concentrations near Pipestone Pond. Willis (1901) made an assessment of these deposits for R.B., and W.D. Job. His report included geology and mineral economics and a proposal that a railway be built to Bay d'Espoir so that ore could be shipped. Moore (1930) reinvestigated the chromite deposits for the Buchans Mining Company and was less enthusiastic than Willis, concluding that they were too 'small and far apart' to be of economic importance.

Snelgrove (1934) described the chromite deposits in some detail as part of a survey of Newfoundland chromite deposits for the Newfoundland Geological Survey. He provided geological sketches of the old trenches, and chemical analyses and petrographic descriptions of the chromite-bearing ultramafic rocks.

During the early 1950's, the Buchans Mining Company carried out extensive base-metal exploration on the original Anglo-Newfoundland Development Company (A.N.D.) charter, including all except the southwest corner of the present area. They produced the first comprehensive geological maps of the area at a scale of $1" = 20,000'$ (Swanson, 1952-1960), which outlined the main geological features. During this program, two detailed mapping projects

were sponsored, which subsequently became university dissertations. Wolofsky (1951) mapped the Sitdown-Pipestone Pond area and described the metasedimentary and metavolcanic rocks and the mafic-ultramafic complex near Pipestone Pond, which he considered to be intrusive. Simultaneously, Slipp (1952) mapped the sedimentary and volcanoclastic sequences in the southeastern part of the area north of Round and Godaleich ponds. The Buchans Mining Company exploration programs in this area were spurred by the 1951 discovery of the South Pond copper deposit. The initial discovery resulted from mapping and boulder tracing, subsequent drilling between 1951 and 1953 (eleven holes totalling 1,760 m) encountered several isolated lenses of chalcopyrite (Buchans Mining Company Limited, 1951-1953). In 1957, Aerophysics of Canada conducted aeromagnetic surveys of the Buchans Mining Company properties (Aerophysics of Canada Limited, 1957) and followed this up in 1958 by geological mapping, soil geochemistry, ground EM and boulder tracing (Pouliot, 1958). The South Pond prospect showed up on ground EM, although it had not been detected by the airborne EM. In 1964, reconnaissance geochemistry was carried out in the volcanic belt along strike from the South Pond prospect (American Smelting and Refining Company Limited, 1965). This work indicated one anomalous zone east of Great Burnt Lake; drilling in 1966 (12 holes) resulted in a chalcopyrite discovery in mafic volcanic rocks. The South Pond mining lease, covering the two deposits and favourable stratigraphy beyond them, was granted effective December 31, 1967 (Figure 1-2). Further drilling of this and the South Pond prospect in 1968, 1969, 1970, 1971 and 1979 together with detailed geochemical, self-potential, magnetic and electromagnetic surveys have outlined significant tonnages of copper-bearing rocks at the Great

Burnt Lake prospect, and at the smaller South Pond prospect. Beavan (1972) summarized previous work and assessed the economic viability of these deposits for Brinex Ltd. He interpreted the Great Burnt Lake deposit as a syngenetic massive sulphide deposit and estimated reserves to be 886,185 tons (803,933 tonnes) averaging 2.47 percent Cu. He noted that the South Pond deposit was generally similar to the Great Burnt Lake deposit and estimated reserves to be 323,000 tons (293,021 tonnes) grading 1.33 percent Cu. Douglas (1976) quoted an estimated tonnage of 1,000,000 tonnes of 3.1 percent copper undiluted for the Great Burnt Lake deposit, but did not give the source of these data. The results of more recent drilling suggest that Beavan's (1972) estimate is probably more realistic with respect to both tonnage and grade.

The chromite occurrences near Pipestone Pond received renewed attention when Fogwill (1964a,b) visited the area as part of a survey of chromite in Newfoundland for the Newfoundland Department of Mines, Agriculture and Resources. He remapped and described the occurrences, provided geological sketches of the prospects, and geochemical analyses of chromite-bearing specimens.

The Hansa Syndicate, holding ground in the volcanic belt south of Great Burnt Lake, did combined airborne EM and magnetometer surveys in 1967 (Sander and Owen, 1968) followed by ground EM and magnetometer surveys and soil geochemistry in 1968 (Sander, 1969). These showed that moderate copper and zinc geochemical anomalies extended from the South Pond Mining lease into the Hansa concession. More detailed work was done in 1969 and 1970 (Sander, 1969; Sander and Heshka, 1970) with EM surveys, copper and zinc soil geochemistry and diamond drilling. Fourteen holes totalling 254 m were drilled on EM anomalies; all were found to have been caused by graphitic horizons.

Coincident with Hansa's work south of Great Burnt Lake, McIntyre Porcupine Mines Limited explored an option from NALCO east and north of the South Pond mining lease. In 1967, they conducted an airborne EM survey (Questor Surveys Limited, 1967), and followed this in 1968 with geochemistry and seven diamond drill holes (843 m), all of which cut graphitic sedimentary rocks in the volcanic sequence (Fox, 1968).

The first published geological map of the whole area was that of Williams (1970) at a scale of 1:250,000. A more detailed map of the mafic-ultramafic complex near Pipestone Pond was published by Kean (1974), who visited the area as part of a survey of the metallogeny of eastern Newfoundland ultramafic rocks for the Newfoundland Department of Mines and Energy. He was the first to suggest that this complex seemed to contain an ophiolitic stratigraphy, and proposed that it could represent a fault-bounded slice of ocean floor rather than an intrusive body. Chemical analyses of chrome spinels collected by Kean were subsequently published by Malpas and Strong (1975).

In 1975, Consolidated Morrison Exploration staked an EM conductor northwest of Godaleich Pond that had been

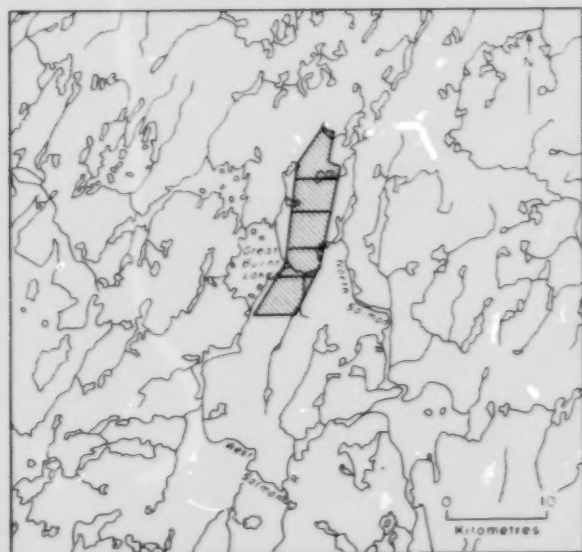


Figure 1-2: Location of the South Pond Mining Lease (cross-hatched area).

found by the Hansa Syndicate on their original concession. Ground EM, magnetic, geochemical and geological surveys were done on the anomaly during 1976 as a joint venture with Riocanex (A. Harris, 1976; V. Harris, 1976; Bucknell *et al.*, 1976). Riocanex did further geological, VLF, magnetic, soil geochemistry and EM surveys in 1977; they found anomalous copper, lead and zinc values in soils and drilled two diamond drill holes (36 m) into graphitic horizons (Bucknell, 1977). Two other claim blocks staked by Consolidated Morrison, northwest and northeast of Cold Spring Pond on Hansa's EM anomalies were investigated by Riocanex in 1976 (Bucknell and McKenzie, 1977a,b). In particular, ground EM and magnetic surveys were done and two holes were drilled in each block (total 634 m), but the anomalies were attributed to graphite.

More recent work in this area by Riocanex concentrated on an area farther to the east. This work included geological mapping, geophysics, geochemistry and diamond drilling (Bucknell, 1979; Harris, 1980a,b). No significant mineralization was reported as a result of this work.

Hudson's Bay Oil and Gas Company carried out limited geological work during 1977 near Great Rattling Brook in the Miguels Lake (2D/12) map area, which led to a find of a mineralized boulder assaying 0.78 percent copper (Dean, 1977a). Subsequent work has been reported by Lassila (1979a,b,c,d).

Lake sediment samples were collected on 4.6-km centres during 1977 as part of a regional lake sediment survey carried out by the Newfoundland Department of Mines and Energy. Multi-element analyses of these samples were reported by Butler and Davenport (1978).

PRESENT INVESTIGATION

This report results from field work carried out during the summer of 1980 by a mapping party under the direction of the author. This work constituted the final phase of a three year study of the regional metallogeny of the Hermitage Flexure; results of the first two years' work have been presented elsewhere (Swinden, 1980a,b; 1981a, 1982). The study was extended into the Pipestone Pond area with the primary aim of identifying the environment of deposition of the South Pond and Great Burnt Lake copper deposits, evaluating the potential for further similar discoveries, and integrating this metallogenically significant part of southern Newfoundland into a regional synthesis. Of equal interest was an evaluation of the massive sulphide potential of the felsic volcanic rocks of the Baie d'Espoir Group, which were known to be present from concurrent mapping in the adjacent Burnt Hill (2D/5) map area (Colman-Sadd, 1981, 1985a) and of possible ophiolitic volcanic rocks of the mafic-ultramafic complex near Pipestone Pond. It was also anticipated that by extending the project into this area, a preliminary assessment could be made of the metallogenic relationship between the Hermitage Flexure region and the time-equivalent volcano-sedimentary sequences of central Newfoundland.

Mapping in the Pipestone Pond area was carried out by two mapping parties from a base camp on the West Salmon River and several aircraft-supported fly camps north and east of the North Salmon River. Traverses were run at an average spacing of 500 m and the data plotted at a scale of 1:20,000. The igneous rock units were systematically sampled for geochemical analysis. Detailed mapping was carried out in the vicinity of known mineral prospects, and the core from several diamond drill holes from the Great Burnt Lake and South Pond prospects was relogged.

Preliminary geological maps at a scale of 1:20,000 have been previously released by the Mineral Development Division (Swinden, 1981b). Descriptions of the rock units and preliminary conclusions concerning their economic potential based on field data were published by Swinden and Collins (1982). The Pipestone Pond area was included in an account of the geology and mineral potential of south-central Newfoundland by Colman-Sadd and Swinden (1982), which incorporated field data collected during the present program. This report contains the most comprehensive and current synthesis of past work and present knowledge in the region as a whole. A more detailed structural history for the area with the implications for the structural history of the Central Mobile Belt was proposed by Colman-Sadd and Swinden (1984a).

Open-file geological maps of the Cold Spring Pond (12A/1) and Great Burnt Lake (12A/8) map areas (Colman-Sadd and Swinden, 1983, 1984b) at a scale of 1:50,000 incorporate data collected during the present program.

PHYSIOGRAPHY

The Pipestone Pond area lies within the Central Plateau of Newfoundland (Twenhofel and MacClintock, 1940). Topography within the area is generally characterized by low, gentle, forested hills which rise from a base elevation of approximately 225 m to between 275 m and 310 m. The highest point in the area lies at 374.2 m on a large hill east of Great Burnt Lake and north of the North Salmon River. Greater than average topographic relief is provided by the Jamieson Hills southeast of Pipestone Pond, which rise sharply to over 325 m and by a large, oval-shaped hill southeast of Great Burnt Lake, which rises to over 370 m.

Drainage within the area is largely controlled by the geological and structural grain. Two prominent rivers drain over 90 percent of the area. The North Salmon River, prior to construction of the North Salmon Dam, was characterized by long, gentle steadies separated by relatively short, fast-flowing and turbulent rapids. Most of the northern two-thirds of the area drains into this river through the Gulp Pond-Pipestone Pond and Sitdown Pond watersheds. The West Salmon River is the focus for drainage in the southern third of the area, and since construction of the Upper Salmon hydroelectric dam, also receives water from the Great Burnt Lake watershed. Both rivers drain into Round Pond and thence into Baie d'Espoir 50 km to the southeast. The northeastern corner of the map area north and east of Gulp Pond drains northwards into the Great Rattling Brook watershed.

PIPESTONE POND AREA

The area has been extensively glaciated and is generally covered by a thin mantle of drift. Large erratics consisting

of foliated granite and granitic gneiss are especially prominent in the western part of the area.

CHAPTER 2

GENERAL GEOLOGY

GENERAL STATEMENT

The Pipestone Pond area lies within the southeastern quadrant of the Newfoundland Central Mobile Belt (Williams, 1964). It forms part of the Dunnage Zone (Williams, 1978) or Terrane (Williams and Hatcher, 1983); the geological units of the Dunnage Zone record the early Paleozoic opening and subsequent closing of the Iapetus Ocean. Within the area, three fault-bounded geological domains exhibit contrasting depositional and/or structural and metamorphic histories (Figure 2-1; Table 2-1). These are, from east to west: 1) quartzite and semipelitic schist assigned to the Spruce Brook Formation (the western part of the Mount Cormack Terrane of Colman-Sadd and Swinden, (1984a), and herein termed the Mount Cormack domain); 2) an oceanic plutonic, volcanic

and sedimentary domain containing ophiolitic sequences of the Pipestone Pond Complex and mafic to felsic volcanic rocks and associated sediments assigned to the Baie d'Espoir Group and the Cold Spring Pond Formation; and 3) psammitic to semipelitic schist and paragneiss intruded by deformed, commonly porphyritic granite and granodiorite, and locally by undeformed equigranular to porphyritic granite (herein termed the western igneous-metamorphic domain).

The easternmost geological domain is underlain by quartzite, mature sandstone and semipelitic schist. The rocks display a progressive eastward increase in metamorphic grade from lower to middle greenschist facies in the Sitdown Pond area to lower amphibolite facies near the eastern edge of the map area. These sedimentary rocks are on strike with, and lithologically similar to rocks in adjacent map areas assigned to the Spruce Brook Formation by Colman-Sadd (1985a). They are intruded locally in the eastern part of the area by granitic pegmatites which are probably related to the Through Hill Granite in the Burnt Hill map area (Colman-Sadd, 1985a).

The Pipestone Pond Complex consists of a westward-facing stratigraphic succession of mafic and ultramafic rocks typical of ophiolitic sequences. The stratigraphically lowest rock units are harzburgite, serpentinized pyroxenite and minor dunite that pass westwards into pyroxenite, gabbro with lesser amounts of diabase and plagiogranite, and mafic volcanic rocks. The complex is disrupted internally by faulting, so that no single stratigraphic section contains the complete ophiolitic stratigraphy.

Volcanic and sedimentary rocks assigned to the Baie d'Espoir Group are present west of Pipestone Pond and south of the North Salmon River. They comprise principally a succession of volcanoclastic rocks, that in some cases display features typical of turbidites, and in others they consist of unbedded, sandstones and conglomerates. A prominent felsic volcanic member is also present. These rocks are on strike and continuous with the North Steady Pond and Salmon River Dam formations in the Twillick Brook map area to the east (Colman-Sadd, 1980a).

A belt of green to grey psammitic and pelitic sedimentary rocks containing interbedded pillow basalts and felsic crystal tuffs outcrops in a northeast-trending belt, which has been traced from west of Cold Spring Pond to the area northwest of Pipestone Pond. This unit, termed the Cold Spring Pond Formation, is assigned formational status later in this report. It hosts two volcanogenic copper deposits north of the North Salmon River. To the east, this formation is faulted against the Baie d'Espoir Group.

The oldest rocks in the western igneous-metamorphic domain are intensely deformed metasedimentary and metavolcanic rocks, which comprise the eastern limits of an

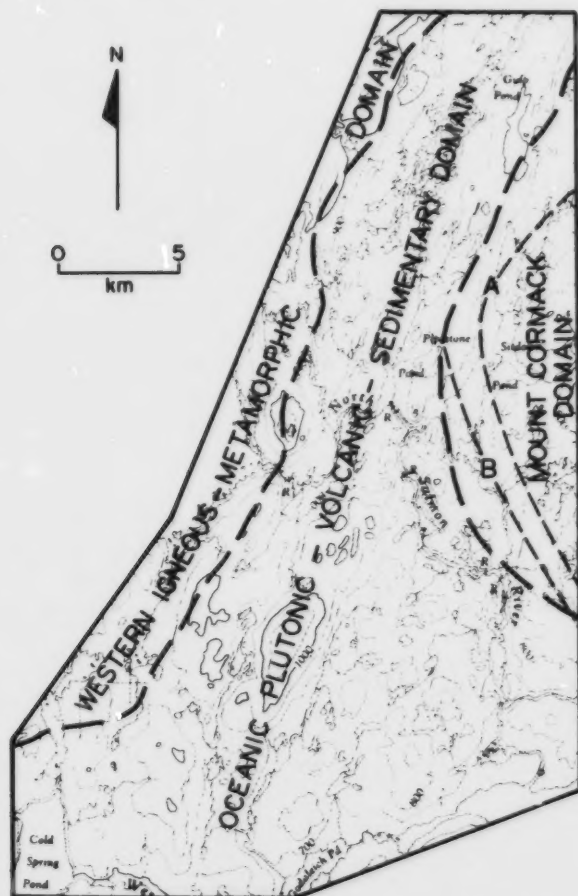


Figure 2-1: Geological domains in the Pipestone Pond area. Heavy dashed lines are boundary thrust faults. Metamorphic isograds in the Mount Cormack domain indicated by light dashed lines. A—andalusite; B—biotite.

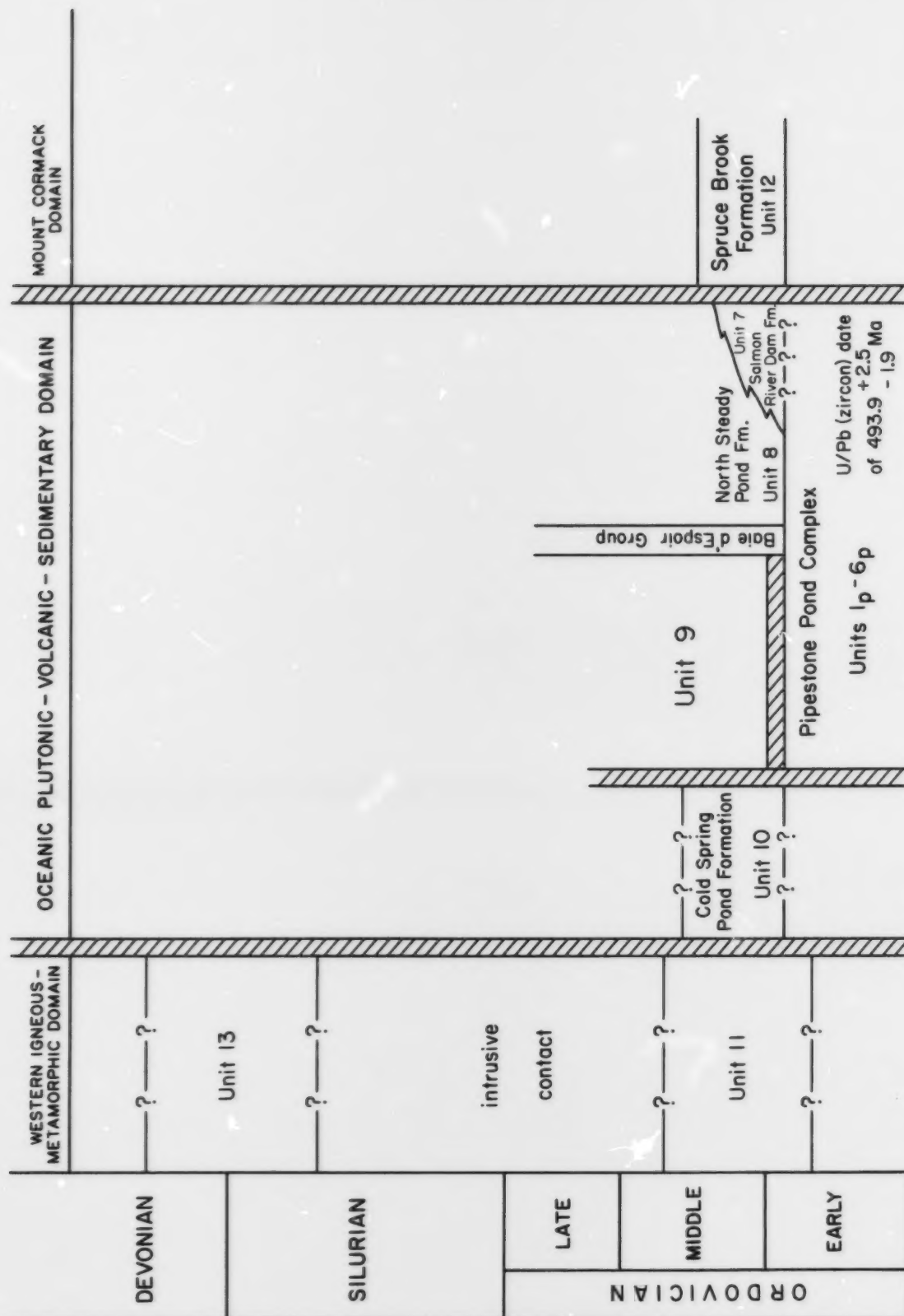


Table 2-1: Table of formations in the Pipestone Pond area.

extensive area of the south-central part of the Dunnage Zone underlain mainly by igneous and metamorphic rocks (see Hibbard, 1983). These rocks are intruded by deformed, porphyritic granitoid rocks, which are locally gneissic and which contain intensely flattened xenoliths of the metasediments. Undeformed biotite, biotite-muscovite granite and granodiorite are represented by numerous boulders near the western limit of the present map area, but are rarely seen in outcrop and are probably derived from areas west of the present map area where similar lithologies have been mapped by Colman-Sadd (1985b).

Some conclusions are possible regarding selective ages of stratified rocks in the different structural domains. The Pipestone Pond Complex is the oldest unit based on a U-Pb (zircon) date of $493.9 \pm 2.5-1.9$ Ma (Dunning and Krogh, 1985). Units 7 to 9 are stratigraphically above the ophiolite. Stratified rocks of Unit 11 are probably in part equivalent to Units 7 to 9, based on lithological similarity. Unit 12 contains Llanvirn fossils and may be the youngest stratified rock unit in the area.

OPHIOLITIC ROCKS (UNITS 1-6)

The oldest rocks in the Pipestone Pond area are ultramafic and mafic rocks of presumed ophiolitic affinity (i.e., derived from an ophiolitic stratigraphic sequence as discussed, for example, by Coleman, 1977). The most extensive and continuous outcrop of these lithologies, extends in a north-south trending, crescent-shaped belt across the eastern half of the area between Gulp Pond and the North Salmon River (Figure 2.2). This unit is herein formally defined as the Pipestone Pond Complex; lithological units of this complex are designated in this report by the subscript 'p'. Numerous smaller occurrences of mafic and ultramafic rocks which are presumed to be of ophiolitic affinity, are preserved along fault traces throughout the Cold Spring Pond Formation and the Baie d'Espoir Group. In the following sections, lithologies within the Pipestone Pond Complex will be described first, followed by the various ophiolitic outliers within the Baie d'Espoir Group and the Cold Spring Pond Formation.

Pipestone Pond Complex (Units 1_p-6_p)

Definition and Distribution

Mafic, ultramafic and minor sedimentary rocks displaying a westward-facing, ophiolite stratigraphy outcrop in a crescent-shaped belt extending from the headwaters of North Great Rattling Brook southward to the area southeast of the Jamieson Hills (Figure 2-2). They are part of a belt of ophiolitic rocks that outcrops from Pipestone Pond eastward to the area of the Northwest Gander River and includes the Great Bear and Coy Pond complexes as well as several smaller unnamed, fault-bounded slivers. Kean (1974) informally referred to this package of rocks as the 'Pipestone Pond ultramafic body', whereas Kean and Jayasinghe (1980), mapping in the Noel Paul's Brook (12A/9) map area, termed

it the 'Pipestone Pond ultramafic complex'. Swinden and Collins (1982), in their preliminary report of the geology of the present area informally referred to it as the 'Pipestone Pond complex', as did Colman-Sadd and Russell (1982) and Colman-Sadd and Swinden (1982 and 1984a). It is herein formally proposed that the name Pipestone Pond Complex be adopted to refer to the package of ophiolitic rocks, including peridotite, pyroxenite, gabbro, diabase, plagiogranite and basalt that outcrop continuously from the area approximately 6 km northeast of Atlantic Lake in the north (Colman-Sadd and Russell, 1982), southward to its termination approximately 2 km southeast of the junction of the North Salmon River and the southward-flowing stream that drains Sitdown Pond. The name is taken from a large pond which is situated near the centre of the complex and on the shores of which most lithologies in the complex can be seen in outcrop. The complex has an exposed strike length of approximately 28 km and attains a maximum outcrop width of approximately 5 km, north of Pipestone Pond.

Six lithological subdivisions are recognized within the Pipestone Pond Complex. No single stratigraphic section exposes all six lithological units. These are described and their type areas designated in the following discussion. An idealized stratigraphic section through the ophiolite is shown in Figure 2-3.

Description of Units

Unit 1_p. Unit 1_p consists of massive, black, fine grained peridotite, including harzburgite and lesser dunite. It forms the eastern boundary of the ophiolitic sequence between Chrome Pond and the northeastern boundary of the map area, and is also exposed in isolated outcrops along the southeastern shores of Pipestone Pond. It is particularly well exposed in a series of low, hummocky hills approximately 2 km east of Gulp Pond. The massive peridotite weathers a characteristic dark-brown colour and its distribution is easily mapped on coloured aerial photographs. Local shear zones contain magnesite-talc alteration. Widely spaced (5 cm) fractures commonly have alteration (serpentinite) haloes up to 5 mm wide and chrysotile cross-fibre (up to 1 mm) occurs locally within them. This unit attains a maximum thickness of 600 to 700 m northeast of Gulp Pond.

A characteristic macroscopic feature of Unit 1_p is the ubiquitous presence of chromite grains ranging up to 3 mm in diameter and commonly forming from 0.5 percent to 2 percent of the rock. In most exposures, minor amounts of chromite appear randomly distributed but in the more chromite-rich exposures, the grains are aligned in parallel bands. Locally, chromite forms blebs up to 10 mm long and in rare cases conformable pods up to 3 m long. A second characteristic megascopic feature of the peridotite is the presence of between 0.5 percent and 5 percent orthopyroxene phenocrysts which range up to 3 mm in diameter.

Peridotite that outcrops adjacent to the eastern boundary fault of the ophiolite north of Chrome Pond and the west boundary fault along the west shore of Pipestone Pond is

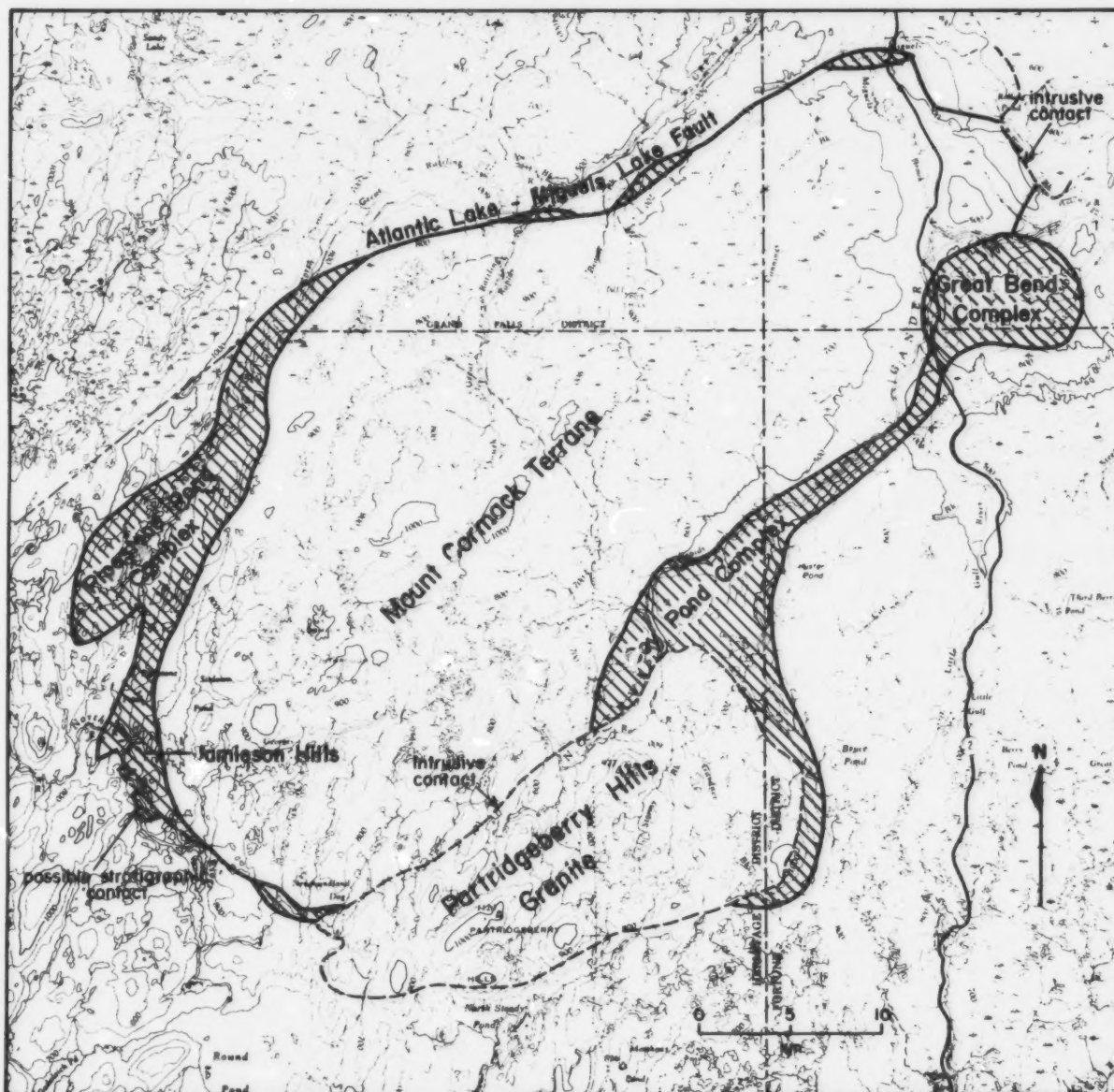


Figure 2-2: Distribution of ophiolitic rocks (represented by hatched diagonal pattern) in the Through Hill area; modified after Colman-Sadd and Swinden (1982).

intensely sheared and altered in a zone generally less than 500 m in width, but ranging up to 900 m wide (Unit 1_{pa}). Outcrops in these sheared zones are bright red on weathered surfaces, a feature which is characteristic of magnesitized outcrops throughout the complex. Structures consisting of elongate, relatively coherent, pillow-shaped blocks up to 2 m long surrounded by an intensely sheared, often comminuted matrix are locally developed. Talc - carbonate schist is well developed west of Chrome Pond but does not appear to be as widespread as in the Coy Pond Complex to

the east (Colman-Sadd, 1985a). In a number of exposures, myriads of quartz veins and veinlets cut the sheared peridotite (Plate 2-1); these are characteristically developed very close to the boundary faults and are particularly well exposed in outcrops east of the Pipestone Pond narrows. Likewise, green mica is locally developed both on fracture surfaces and associated with quartz veins.

In thin section, the peridotite is commonly seen to be pervasively serpentinized. Fibrous antigorite is the most

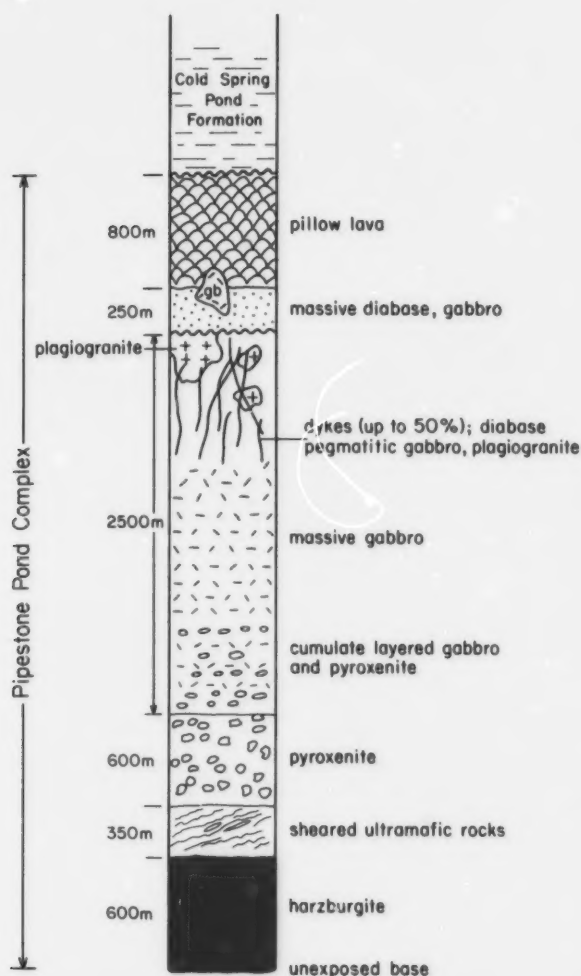


Figure 2-3: Schematic idealized stratigraphic section through the Pipestone Pond Complex. Thicknesses are based on exposed maximum widths of units and do not account for effects of folding or intraformational faulting. Wavy contacts are faults, straight-line contacts are stratigraphic.

common alteration product and in some samples, it comprises 90 percent of the rock, leaving no trace of the original mineralogy. In other cases, alteration has been less complete and relict olivine and pyroxene crystals are preserved in a dominantly antigorite matrix (Plate 2-2). In a few relatively unaltered specimens, olivine, clinopyroxene and orthopyroxene are the principal rock-forming minerals in approximate proportions 65:30:5. Olivine in these samples forms subrounded grains not uncommonly altered to serpentine on grain rims and internal fracture surfaces. The principal minor component is chromite which generally comprises less than 3 percent.

The intensely tectonized samples from Unit 1_{pa} are in all cases rich in magnesite as well as antigorite and seldom

retain more than isolated fragments of original mineralogy (except for chromite). Magnesite along with talc, forms a fine grained, comminuted matrix to brecciated larger fragments of serpentinite and magnesite in some samples. In other cases, magnesite forms large (up to 5 mm) euhedral porphyroblasts which overgrow the serpentinite fibres and may comprise up to 60 percent of the rock (Plate 2-2). Quartz veining, where present, appears to postdate magnesitization and serpentinization, which appear to be approximately contemporaneous.



Plate 2-1: Peridotite outcrop near Chrome Pond cut by numerous quartz veins.

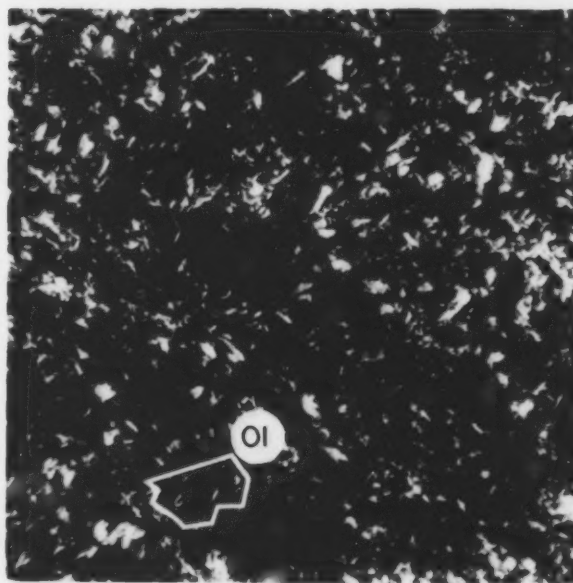


Plate 2-2: Photomicrograph of magnesitized peridotite. Fibrous mineral is antigorite. Olivine remnant in lower left (ol). Field of view is 6 mm.

Unit 2_p. Unit 2_p consists of banded peridotite and pyroxenite, which is generally less than 400 m thick. In the northern part of the complex, it occupies a laterally continuous belt between the peridotite (Unit 1) and pyroxenite (Unit 3) and appears to form a transitional unit from one to the other. The eastern contact between this unit and the peridotite appears to be abrupt and may be faulted, although the actual contact was not observed on the ground. The transition is marked by the reddening of weathered, peridotite outcrops and the subsequent appearance of green-weathering pyroxenite bands. Easterly outcrops of Unit 2_p comprise mainly fine grained, red-weathering, altered, peridotite containing disseminated chromite grains, but lacking the orthopyroxene phenocrysts characteristic of Unit 1. Minor shearing subparallel to the Pipestone Pond Complex boundary faults is evident, but the rocks are not sheared to the extent of those red peridotites that make up Unit 1_{pa} near the boundary faults. Local shear zones and minor development of talc-carbonate schist in this unit may be related to movement on local, internal faults. Progressing towards the west and generally within 100 m of the transition to red-weathering peridotite, medium- to coarse-grained pyroxenite, occurring as pods, lenses and as parallel bands of from 2 cm to several metres in thickness becomes more prevalent. Some very coarse grained bands contain individual pyroxene crystals up to 4 cm across (Plate 2-3). The relationships between the peridotite and the pyroxenite are not always clear. In some areas, pyroxenite veinlets intrude peridotite, whereas in others, there is a sequential, apparently stratigraphic banding of the two lithologies. The western boundary of Unit 2_p in the northern part of the complex is gradational with the pyroxenites of Unit 3; as a general rule, the pyroxenite/peridotite ratio progressively increases westward until the latter lithology disappears. The most westward exposure of

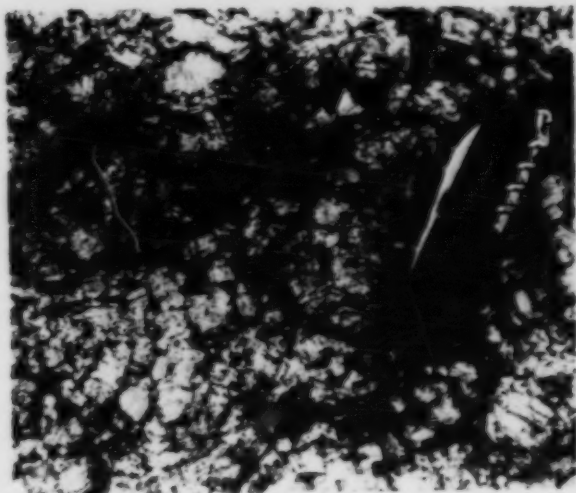


Plate 2-3: Very coarse grained pyroxenite from northeast of Pipestone Pond.

red-weathering peridotite is taken as the boundary between Units 2_p and 3_p.

In the southern part of the Pipestone Pond Complex, Unit 2_p forms the stratigraphically lowest exposed unit. It is faulted to the east against metasedimentary rocks of Unit 12 and to the west against gabbro of Unit 4_p. The best exposures of Unit 2_p in the whole complex are found in this area, especially in the east-central Jamieson Hills. Pyroxenite-peridotite banding is evidenced on a much broader scale in this area, and bands up to several tens of metres wide can be traced for hundreds of metres along the strike. In some cases, these individual bands are large enough to be mapped as individual units at a scale of 1:20,000. The lithologies present here are otherwise similar to those found in the northern part of the complex.

In thin section, the peridotites of Unit 2_p are seen to be almost completely altered to fibrous serpentine (antigorite) and magnesite. Relict olivine grains, now completely serpentinized, are seen in a few places. Magnesite, which generally forms 30 to 60 percent of the rock, occurs as fine grained masses, in veinlets and in relict olivine crystals either intergrown with serpentine or as rhombohedral grains overgrowing the serpentine. Pyroxenite bands consist largely of an interlocking mosaic of clinopyroxene with up to 15 percent finer grained interstitial olivine and orthopyroxene. The clinopyroxenes appear relatively unaltered although chlorite is locally developed at grain margins. Commonly, the olivine and orthopyroxene are partially to completely replaced by a mosaic of serpentine and magnesite, locally accompanied by opaque minerals. Serpentine and magnesite veinlets cut many olivine crystals and extend along grain boundaries and fracture surfaces into the adjacent pyroxenes. Chromite appears to be absent from the pyroxenites.

Unit 3_p. Unit 3_p consists dominantly of medium- to coarse-grained pyroxenite and layered pyroxenite and gabbro. It forms a transitional zone up to 700 m thick between the ultramafic rocks (Units 1_p and 2_p) and the mafic rocks to the west (Unit 4_p). Unit 3_p was recognized only in the northern part of the complex where it outcrops in a continuous belt from 200 to 700 m wide between the north end of Pipestone Pond and the area north of Gulp Pond; in the southern part of the complex, it is apparently faulted out of the succession.

Pyroxenite without gabbro forms a relatively narrow belt along the eastern boundary of Unit 3_p, east of the last exposures of serpentinized peridotite. Small layers of gabbro first appear in the lower part of the section and gradually increase in abundance and size towards the west. Gabbro and pyroxenite are generally layered on outcrop scale in the western part of Unit 3_p. Dykes of the fine grained gabbro and diabase also cut the pyroxenite in this area. The western contact with Unit 4_p is drawn to reflect the point at which pyroxenite becomes subordinate to gabbro in the section.

Pyroxenites of Unit 3_p tend to be more feldspathic than their equivalents in Unit 2_p to the east and also to be finer grained, with individual crystals ranging up to 2 mm. Otherwise, both pyroxenite sequences appear similar in outcrop. Likewise, gabbro bands in Unit 3_p are lithologically and petrographically similar to those of Unit 4_p (see below).

Unit 4_p. Unit 4_p consists of a suite of mafic and felsic intrusive rocks, which comprises mainly equigranular, medium- to coarse-grained pyroxene gabbro, medium grained diabase, pegmatitic gabbro and plagiogranite. This unit is more extensively developed in the northern part of the complex where it outcrops in a large, elongate lens extending from the area northwest of Pipestone Pond to the north end of Gulp Pond. Unit 4_p also outcrops in the southern part of the complex where it forms a narrow outcrop band in the Jamieson Hills east of Pipestone Pond and the North Salmon River.

In the northern gabbro lens, the principal lithology is medium- to coarse-grained gabbro which constitutes at least 70 percent of the unit. In the eastern (stratigraphically lower) areas, the gabbro is interbanded with medium- to coarse-grained pyroxenite (similar to Unit 3_p), both on a mesoscopic and microscopic scales; the amount of pyroxenite gradually decreases towards the west until it disappears altogether. As previously noted, the eastern contact between Units 3_p and 4_p is taken as the approximate point where gabbro becomes predominant over pyroxenite in the section. The western (upper) part of the body is faulted against sedimentary and volcanic rocks of the Baie d'Espoir Group and the Cold Spring Pond Formation.

Westerly exposures of gabbro are commonly cut by dykes of diabase, pegmatitic gabbro and plagiogranite. The diabase dykes generally occur in swarms and commonly are injected along parallel tracks, which maintain a uniform spacing of 20 cm to 1 m and comprise 35 to 50 percent of some outcrops (Plate 2-4). They are generally not sheeted in the sense that dyke-within-dyke features cannot be recognized; however, parallel dykes are locally seen to share a mutual contact along which screens of gabbro are preserved. The dykes range up to 1 m in width.

Pegmatitic pyroxene gabbro (Plate 2-5) is a common constituent throughout the western part of this body and is particularly well developed southwest of South Pond where it forms approximately 5 percent of the exposure. It consists of very coarse plagioclase and pyroxene crystals (up to 10 cm), and locally contains blebs of magnetite up to 2 cm in diameter. It also occurs as dykes ranging from 30 cm to greater than outcrop width, which cut gabbro and diabase and are in turn locally cut by diabase and plagiogranite. At some locations, pegmatitic gabbro comprises entire outcrops up to 10 m across.

A third intrusive phase which is common in the upper part of the western gabbro body is plagiogranite, which generally forms dykes and lenses within the gabbro. It both



Plate 2-4: *Diabase dyke swarm intruding coarse grained, equigranular granite.*

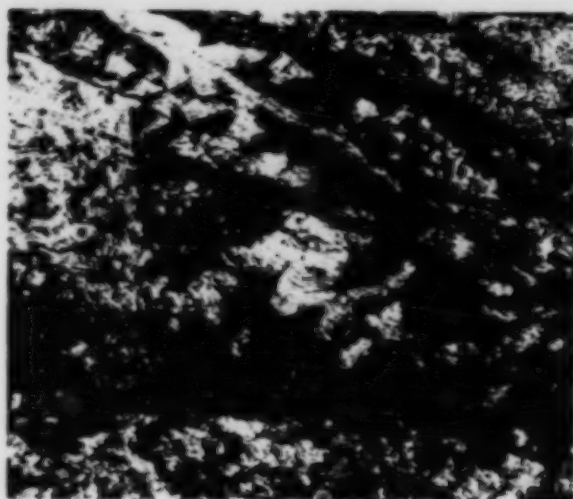


Plate 2-5: *Very coarse grained pegmatitic pyroxene gabbro. Light-coloured mineral is plagioclase, dark is clinopyroxene. Single crystals are up to 10 cm long.*

cuts, and is cut by fine grained diabase dykes and pegmatitic gabbro. Plagiogranite locally forms large, irregular pods which are separable into a distinct map unit at 1:50,000 scale (Unit 5_p). It also occurs as complex net veining within outcrops predominantly composed of gabbro.

In the gabbro bodies south and east of Pipestone Pond, fine- to medium-grained mafic intrusive phases predominate, such that much of this outcrop would probably be more properly termed diabase than gabbro. Many of the outcrops, particularly those close to the contact with the pillow lava,

consist of very fine grained green rocks with little or no texture or structure recognizable in outcrop or hand specimen. Surfaces are deeply weathered and lichen covered so that sheeted dykes could possibly pass unrecognized. Somewhat coarser phases underlie the northern extremity of this body west of Chrome Pond and in this area they are commonly cut by plagiogranite and diabase dykes. Lesser amounts of plagiogranite also cut the finer grained phases in the Jamieson Hills. There is also a small area underlain by banded gabbro-pyroxenite near the shore of Pipestone Pond southeast of the narrows. Unit 4_p in this area is faulted to the east and north against peridotite and serpentinite of Units 1_p and 2_p. The only exposed gabbro-pillow lava contact occurs on a hill approximately 300m southwest of the south end of Chrome Pond. At this locality, the contact between medium-grained gabbro and the volcanic rocks appears to be intrusive. Here sheeted dykes are not present between the gabbro and volcanic members of the ophiolite.

In thin section, most medium- and coarse-grained gabbros and pegmatitic gabbros are seen to consist principally of clinopyroxene and plagioclase. The clinopyroxenes are moderately to strongly altered to antigorite-talc-actinolite, but are rarely strongly uralitized. Relatively fresh grains commonly show twinning and/or exsolution lamellae. Orthopyroxene is a minor component in some samples, occurring as relatively small interstitial grains, often highly altered. Olivine is less commonly seen, occurring likewise as small, highly altered interstitial grains. Green amphibole is locally present, generally as small euhedral crystals and as replacement patches in large pyroxene grains.

Plagioclase comprises from 25 to 50 percent of most specimens and is always strongly saussuritized with the development of chlorite-zoisite-epidote-actinolite assemblages. Original features of the feldspar grains have been destroyed by the alteration, hence determination of feldspar composition was impossible. Magnetite is a common accessory.

In thin section, finer grained phases of the gabbro bodies (e.g., diabase dykes and fine grained gabbros) generally consist of a granular mass of intergrown, saussuritized, equant plagioclase crystals with a mafic matrix comprising green hornblende, chlorite, actinolite and uralitized pyroxene.

The stratigraphic relationships between gabbros in the northern and southern parts of the complex are not entirely clear. The former contains the stratigraphic base of the gabbro, but because its upper contact is always a fault, it is not known how close this section approaches the top of the unit. The southern gabbro body appears to represent the stratigraphic top of the gabbro as its upper contact with the pillow lavas is exposed in this area. However, the lower contact is always faulted. The range of overlap or size of stratigraphic discontinuity between the gabbros in these two areas is not known. However, gabbro lithologies in the two areas are very similar in grain size, range of minor intrusions

and general appearance, suggesting that there is at least some overlap or, at most, a minor discontinuity.

Unit 5_p. Unit 5_p consists of lenses of plagiogranite, which intrude the upper levels of the gabbro in both the northern and southern parts of the complex. Plagiogranite is very common in this part of the stratigraphy, occurring principally as dykes and lenses that are too small to be separated at map scale from the other lithologies of Unit 4_p. However, four larger plagiogranite bodies have been delineated that can be shown at map scale; three in the northern gabbro unit and the other in the southern gabbro unit, east of Pipestone Pond. In hand specimen, the plagiogranite is a leucocratic, medium- to coarse-grained rock composed of quartz, plagioclase and minor chloritized mafic minerals. It is differentiated from the leucocratic gabbro varieties by the ubiquitous presence of prominent quartz crystals, which stand out as knobs on weathered surfaces. Field relations suggest that the plagiogranite is genetically related to the ophiolite as it both cuts and is cut by diabase dykes that are an integral part of the Unit 4_p gabbro bodies.

In thin section, the plagiogranite comprises coarse grained quartz and fewer plagioclase phenocrysts set in a fine grained groundmass. Quartz content is variable, generally comprising 30 to 40 percent of the rock. The phenocrysts are generally prismatic, commonly from 1 to 2 mm in length but ranging up to 5 mm. Most crystals are clear, commonly without subgrain boundaries and are variably strained as evidenced by undulose extinction. Recrystallization along grain boundaries is sometimes seen and crystals are commonly broken by cataclastic deformation.

Plagioclase forms both a phenocryst phase and a fine grained component of the groundmass. Phenocrysts are commonly smaller than quartz, ranging up to 3 mm in length. They are rarely unaltered, and are commonly dusty in appearance containing abundant inclusions of actinolite, epidote and chlorite. Compositions are in the range An₅₅-An₆₅ (labradorite). Myrmekitic texture was noted in one thin section; zoning is locally present, but not common.

The fine grained groundmass is predominately plagioclase with lesser amounts of quartz and mafic minerals, now completely altered to actinolite, epidote ($\pm$ clinozoisite), chlorite and a dusty opaque mineral (probably magnetite). In places, ragged pseudomorphs suggest that the predominant mafic phase was pyroxene. Zircon and sphene are accessories. The groundmass textures are commonly highly cataclastic with broken grains and patches of comminuted material. Cataclastic deformation has locally introduced a strong penetrative foliation.

Unit 6_p. Unit 6_p consists of mafic volcanic rocks and minor interbedded sedimentary rocks, and is preserved only in the southern part of the complex. The volcanic rocks outcrop in a northwest-trending belt, which straddles the North Salmon River and the western side of the Jamieson Hills. In the northern and central Jamieson Hills, the volcanic

rocks directly overlie and are intruded by gabbros of Unit 4_p; south of the Jamieson Hills, they are faulted against ultramafic rocks of Unit 2_p. The western contact between these volcanic rocks and conglomerates of the North Steady Pond Formation (subunit 8c) is interpreted as depositional, and is of considerable interest as a major factor in constraining interpretations of the sedimentary environment of the Baie d'Espoir Group in this area. It is described in more detail in the description of subunit 9c and its significance is discussed in Chapter 6.

The principal rock type in Unit 6_p is basalt consisting mainly of pillow lavas, but including massive flows and fine grained mafic tuffs. The pillow lavas consist of well-developed, generally unflattened pillows from 0.3 to 1.6 m in length. Many pillows have highly vesicular cores and spectacular variolitic textures in the outer regions (Plate 2-6). Pillow selvages are generally chloritized and epidotization is common in both the rims and cores. Calcite and chlorite fill most vesicles and carbonate veining is locally developed. Pillow breccias consist of 50 to 80 percent rounded, vesicular, locally variolitic, pillow fragments (10 to 50 cm in diameter) in a fine grained hyaloclastic matrix, and are locally interbedded with intact pillow lavas. Fine grained tuff and tuffaceous sedimentary rock, commonly consisting of green, highly cleaved (commonly fissile) rocks that display no obvious sedimentary structures apart from rare, poorly preserved bedding planes, are present in a few exposures within the sequence. These rocks are probably fine grained hyaloclastic and/or epiclastic rocks, which have resulted from reworking of the mafic volcanics. All of the above lithologies are well exposed along the shores of the North Salmon River west of the Jamieson Hills.

In thin section, the basalts are seen to be highly altered, retaining very little original texture or mineralogy. Strongly saussuritized plagioclase occurs very rarely as phenocrysts. The groundmass consists of 30 to 40 percent randomly oriented laths, which occasionally retain some relict feldspar, but more commonly are completely altered to chlorite-calcite-epidote. Laths range up to 0.5 mm but are generally less than 0.2 mm. Mafic minerals in the groundmass have been completely altered to actinolite-epidote-chlorite and rarely to hornblende.

Calcite, accompanied by minor quartz veining, is widespread in most sections and calcite inclusions form a prominent part of the alteration assemblage. Amygdules are present in many samples and are commonly filled with calcite and lesser amounts of epidote.

Dark grey to black, fine grained clastic sedimentary rocks outcrop in scattered shoreline exposures along the North Salmon River. The rocks are commonly highly sheared and altered, such that original sedimentary features are seldom preserved. These rocks are highly recessive and it is possible that they are more widespread in the volcanic sequence than the present exposure indicates. The relationship of these rocks



Plate 2-6: *Varioles in pillow basalts of the Pipestone Pond Complex along the North Salmon River.*

to the ophiolite is not well defined. They may be interbedded with the pillow lava sequence and, due to their highly recessive nature, only exposed along the river, which also marks the exposed top of the pillow lavas. Alternatively, they may depositionally overlie the ophiolite and in this event would represent the highest stratigraphic level presently exposed. Similar rocks in the Burnt Hill (2D/5) map area to the east have been interpreted to overlie the Coy Pond Complex, the eastward extension of the Pipestone Pond Complex (Colman-Sadd, 1985a). However, in view of the fact that these sedimentary rocks are highly sheared in most exposures, it is also possible that they have been emplaced along faults and are not genetically related to the ophiolite.

Contact Relations

The ophiolite is almost everywhere considered to be faulted against adjacent rocks. To the east, it is thrust over quartzose sedimentary rocks of the Spruce Brook Formation (Unit 12) along a steeply dipping fault, which has a well-defined topographic expression. In the west, the ophiolite is overthrust by volcanic and sedimentary rocks of the Baie d'Espoir Group and the Cold Spring Pond Formation along a complex fault system that overrides, in various areas, every stratigraphic unit in the ophiolite. The trace of this fault has a strong topographic expression in the area west and north of Gulp Pond and in the area north and northwest of Pipestone Pond, but in other areas the topographic expression is less prominent.

Structure and Metamorphism

The Pipestone Pond Complex is a sequence, which everywhere appears to face westward. There do not appear to be major stratigraphic disruptions or repetitions. Dips within the complex range from approximately 20° to 65° westward; Stuckless (1975) has suggested on geophysical grounds that the complex as a whole dips beneath central Newfoundland at approximately 60°. There is some evidence for relatively minor movement of the internal parts of the ophiolite along thrust faults subparallel to the boundary faults. East of Gulp Pond, a sliver of massive peridotite outcrops west of the banded peridotite-pyroxenite of Unit 2_p, possibly indicating its emplacement along a fault. This is the only recognized example of repetition of the stratigraphy by faulting; however, there are several recognized internal faults, which apparently exclude part of the stratigraphy. One of these faults trends southwards from the southwest corner of Chrome Pond through the Jamieson Hills, juxtaposing banded ultramafic rocks of Unit 2_p with gabbro and volcanic rocks of the Pipestone Pond Complex (Units 4_p and 6_p) and, ultimately, with sedimentary rocks of the Baie d'Espoir Group. Likewise, peridotite of Unit 1_p appears to have been cut out by thrust faulting southeast of Gulp Pond. Intense shear zones of limited extent occur throughout the complex, and probably mark the traces of minor internal faults.

Perhaps the most important stratigraphic disruption within the complex is caused by a transverse, east-northeast-trending fault, which has been traced from the southwest corner of Chrome Pond westward to the shore of Pipestone Pond. This fault divides the complex into two structural blocks. North of the fault, the lower part of the ophiolitic stratigraphy is exposed from the harzburgite layer in the east to gabbro, diabase and plagiogranite in the west. South of the fault, higher stratigraphic levels are exposed; the lowest level present here is banded ultramafic rock of Unit 2_p, which is succeeded by gabbro and, ultimately, volcanic rocks. The volcanics form the highest stratigraphic levels of the ophiolite and are not preserved in the northern block.

A regional tectonic fabric is variably developed in the lower parts of the ophiolite. As previously mentioned, most zones of shearing appear to be of local extent and related to emplacement of the ophiolite in its present structural position.

However, tuffaceous members of the mafic volcanic sequence southeast of Pipestone Pond possess a strong, northeast-to north-northeast-trending cleavage, which is vertical to steeply northwest dipping. This cleavage is not parallel to the boundary faults, but is parallel to similar cleavage developed in the Baie d'Espoir Group to the south and west. It seems likely, then, that the development of this fabric postdates the emplacement of the ophiolite. A faint foliation in the pillow lavas, defined in thin section by growth of fibrous actinolite and chlorite, may be the same fabric as discussed above.

Metamorphism of mafic rocks throughout the complex resulted in the growth of chlorite-epidote ($\pm$ clinozoisite)-actinolite ($\pm$ hornblende) and took place in the lower- to middle-greenschist facies.

Correlation and Age

The Pipestone Pond Complex can be traced well beyond the limits of the Pipestone Pond area (Figure 2-2). As previously noted, its northern end has been traced across the southeast corner of the Noel Paul's Brook map area (Kean and Jayasinghe, 1980) to the southwest corner of the Miguels Lake map area (Colman-Sadd and Russell, 1982), where it is finally terminated between converging boundary faults east of North Great Rattling Brook. The eastern extension of these boundary faults has been traced for more than 20 km across the Miguels Lake sheet where Colman-Sadd and Russell (1982) have referred to it as the Atlantic Lake-Miguels Lake fault zone. Rocks probably related to the Pipestone Pond Complex commonly occur in fault slivers along this fault zone (Colman-Sadd and Russell, 1982). East of the Miguels Lake map area, the fault zone strikes into the younger Mount Peyton intrusive suite. It emerges with a southwesterly trend at the south end of this intrusive suite near Rolling Brook (Figure 2-2), and again widens to expose the Great Bend and Coy Pond complexes, before being cut off by the Partridgeberry Hills Granite (Colman-Sadd, 1981, 1985a; Colman-Sadd and Swinden, 1982). The southern end of the Pipestone Pond Complex is pinched out between converging boundary faults east of the North Salmon River, but the fault zone can be traced southeastward by the occurrence of sparse slivers of ultramafic rocks for another 6 km. This fault zone is ultimately cut off by the southwest corner of the Partridgeberry Hills Granite (Colman-Sadd, 1980a). It is therefore reasonably well established that the Pipestone Pond Complex forms part of a larger ophiolite belt which outcrops in a 65 km-long, elliptical belt (Colman-Sadd and Russell, 1982; Colman-Sadd and Swinden, 1982, 1984a). This belt contains two reasonably complete ophiolite sequences, the Pipestone Pond and Coy Pond Complexes, and a third partial sequence, the Great Bend Complex. These are linked by a continuous fault zone which is broken only by the post-emplacement intrusion of the Partridgeberry Hills Granite and the Mount Peyton intrusive suite.

The ophiolitic rocks exposed in this belt are virtually identical both lithologically and in terms of their structure, metamorphism and tectonic setting to rocks of the Gander River Ultrabasic Belt (GRUB) defined and described by Blackwood (1982). The GRUB has been interpreted to

represent oceanic crust tectonically emplaced on the continental margin east of the Iapetus Ocean (Currie *et al.*, 1979; Blackwood, 1982), and a similar origin was proposed for the Pipestone Pond–Coy Pond ophiolite belt by Colman-Sadd and Swinden (1982, 1984a) who considered the two belts to be equivalent.

Because the Pipestone Pond–Coy Pond ophiolitic belt has not yielded any fossils and its boundaries are almost everywhere faulted, there is no direct stratigraphic evidence of its age. Indirect evidence is afforded by ophiolitic sequences of central and western Newfoundland which are generally considered to be similar in age to the Pipestone Pond Complex (e.g., Dean, 1978). For example, the ophiolitic rocks of western Notre Dame Bay are overlain by strata that contain Arenigian fossils (Snelgrove, 1931; MacLean, 1947; Dean, 1978), and if it is assumed that these are similar in age to the Pipestone Pond Complex (not entirely an unfounded assumption; see Dunning and Krogh, 1985), then this provides a minimum age for generation of the ophiolitic rocks.

The only direct evidence for the absolute age of the Pipestone Pond Complex is the U–Pb (zircon) date of 493.9 ± 2.5 – 1.9 Ma reported by Dunning and Krogh (1985) on a plagiogranite sample from east of Pipestone Pond. This date is slightly older than dates that the same authors obtained from ophiolitic rocks in southern and western Newfoundland (the Betts Cove, Bay of Islands and Annieopsquotch complexes), suggesting that although the various ophiolites may be broadly coeval, they may not be exact time equivalents.

Other Ultramafic–Mafic Rocks of Ophiolitic Derivation

Presumed ophiolitic rocks that are not part of the Pipestone Pond Complex outcrop in small, fault-bounded slices throughout the volcanic and sedimentary terranes west of the ophiolite. (These are designated as Units 1–6 without a subscribed 'p' on the geological map.) All lithologies of the Pipestone Pond Complex are present in these slivers, although nowhere are they all present in a single lens.

The largest of these ophiolitic slices underlies the area immediately north of Cold Spring Pond and is faulted against sedimentary and volcanic rocks of the Cold Spring Pond Formation to the south and east and metasedimentary rocks of Unit 7 to the north. It contains several ophiolitic lithologies and although poorly exposed, appears to consist of at least three fault-bounded lithological units. Peridotite and sheared, serpentized, magnesitized peridotite underlie the northwestern part of the slice. The predominant lithology is very similar to that found east of Gulp Pond (Unit 1p). Red-weathering magnesitized peridotite forms a minor component in the southwestern part of this slice, whereas the northeastern part of the slice is underlain by medium- to coarse-grained gabbro and plagiogranite, which are lithologically similar to those in the upper parts of the gabbro body northwest of Pipestone Pond. The gabbro and ultramafic lithologies are bounded to the south and southeast by sheared and deformed

pillow lava, which is interpreted on the basis of its association with other ophiolitic rocks and its geochemistry (see Chapter 3) as an outlying fragment derived from the Pipestone Pond Complex.

Rocks of presumed ophiolitic affinity outcrop in two small slices northeast of Cold Spring Pond and east of the North Salmon Dam access road. They consist predominantly of medium grained gabbro, diabase and plagiogranite, although the more northerly of the two slices also contains sheared, serpentized peridotite. A third large slice of mainly gabbroic rocks outcrops along the North Salmon River east of the North Salmon Dam.

There is also a relatively large fault-bounded slice of inferred ophiolitic rocks immediately northeast of Pipestone Pond, containing gabbro and plagiogranite at its southern end and pillow lava to the northeast. It is faulted against pyroxenite of the Pipestone Pond Complex to the northeast and elsewhere against sedimentary rocks of the Baie d'Espoir Group. The pillow lavas in this sliver were previously interpreted by Swinden (1981b), Swinden and Collins (1982) and Colman-Sadd and Swinden (1982) to be part of the Baie d'Espoir Group. However, from geochemical evidence (see Chapter 3,) it is suggested that they are more likely related to basalts in the Pipestone Pond Complex and are therefore reassigned to Unit 6.

The only rocks of possible ophiolitic affinity exposed entirely within the outcrop area of the Baie d'Espoir Group are sheared basalts and hyaloclastites in the southeast corner of the map area along the North Salmon River north of Round Pond. No definite pillows were observed here but abundant vesicles in several exposures indicate that the rocks are volcanic. Other outcrops contain quite fissile, dark green, chloritic rocks, which may represent tuffaceous or sedimentary rocks. In thin section, the basalt flows are seen to consist of a fine grained mosaic of plagioclase laths 0.05 to 0.2 mm long within a fine grained, altered groundmass consisting principally of chlorite pseudomorphs of very fine grained pyroxene and amphibole. Some relict pyroxene remains. Apatite, sphene and opaque minerals are accessory phases.

These basalts were originally assigned to the Baie d'Espoir Group (Swinden and Collins, 1982) and interpreted as depositional lenses within the succession. However, geochemical evidence (see Chapter 3) suggests they are better interpreted as related to Pipestone Pond Complex basalt, and they are therefore reassigned to Unit 6.

The structural setting of these ophiolitic slivers is not well defined. Two possibilities exist. It may be that the faults which bound them are major, steep, east- and northeast-trending structures that extend beyond the boundaries of the ultramafic rocks in both directions. In this case, the traces of these faults will be very difficult to recognize as they would continue into the adjacent sedimentary rocks where they would separate similar lithologies. The exposure in the surrounding area is poor so that the faults would not be

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expected to be exposed. Alternatively, the ophiolitic rocks may be exposed in windows formed as a result of the upward doming of underlying strata; in this case, the bounding fault traces would be relatively flat and would not outcrop beyond the margins of the ophiolitic slivers. If this interpretation is correct, it would imply that ophiolitic rocks structurally underlie at a relatively shallow depth all rocks west of Pipestone Pond to the limits of the present mapping.

Very small slivers of predominantly ultramafic rock are present along fault traces in the northern part of the map area, particularly in the area southwest of Pipestone Pond. Single outcrops of sheared peridotite within the Cold Spring Pond Formation and Baie d'Espoir Group volcanic-sedimentary sequences occur along the traces of both major and minor faults, and it appears that the ultramafic rocks have been preferentially mobilized along these faults. Three northeast-trending faults southeast of Pipestone Pond were identified by this feature. In addition, the location of the major fault that juxtaposes the Cold Spring Pond Formation with the igneous-metamorphic domain to the west is marked in at least two locations (on the road southwest of the North Salmon Dam and northwest of Gulp Pond) by sheared peridotite. Although most of the fault-bounded ophiolitic slices occur within the outcrop area of the Baie d'Espoir Group and the Cold Spring Pond Formation, a single large slice of serpentinized peridotite occurs within porphyritic foliated granitoid rocks northeast of Great Burnt Lake. The contacts of this slice are not exposed and it is not clear whether it is a fault-bounded sliver within the granitoid rocks or a large xenolith.

It should be emphasized that although ophiolitic rocks appear on faults in all units west of the Pipestone Pond Complex, none have been discovered in the sedimentary sequences of the Mount Cormack domain east of the complex.

In thin section, the rocks which outcrop in these slices are not significantly different from similar lithologies in the Pipestone Pond Complex. The peridotites are always extensively sheared (with the exception of those in the larger-than-normal slice north of Cold Spring Pond), possibly reflecting concentrated shearing, which they suffered on these faults during emplacement. Otherwise, their mineralogy and textures appear to be the same. They are considered to be parts of the same ophiolitic sheet which is now exposed in the Pipestone Pond-Coy Pond ophiolitic belt and also in the GRUB.

BAIE D'ESPOIR GROUP (UNITS 7-8)

Definition and Subdivision

Jewell (1939) originally defined the Baie d'Espoir Series to include sedimentary rocks (now known to include some volcanic rocks as well) exposed around the coastline of Bay d'Espoir. Anderson (1967) assigned group status to these rocks.

The Baie d'Espoir Group has previously been subdivided into five formations. Four of these, the Salmon River Dam, St. Joseph's Cove, Riches Island and Isle Galet formations were defined by Colman-Sadd (1976) and the fifth, the North Steady Pond Formation, by Colman-Sadd (1980a). Colman-Sadd (1980b) outlined the probable stratigraphic relationships of the various formations of the Baie d'Espoir Group. In this model, the Salmon River Dam Formation was interpreted to be the lowermost unit, representing clastic sediments laid down at the base of the Baie d'Espoir Group depositional basin, and presently exposed where they have been upwarped by the intrusion of later granites. The Salmon River Dam Formation was interpreted to be overlain by the St. Joseph's Cove and North Steady Pond formations. These are in part lateral equivalents, which consist predominantly of clastic sedimentary rocks (accompanied by felsic flows and pyroclastic rocks in the latter) and exhibit abundant evidence of deposition by turbidity currents.

Salmon River Dam Formation (Unit 7)

Definition and Distribution

The Salmon River Dam Formation outcrops in an east-west trending belt across the southern boundary of the present area, and has been traced from the area north of Round Pond westward to the West Salmon River. It forms the southern limit of geological interest with respect to the present program and was not examined in detail. Typical exposures of this formation are found in river sections along the West Salmon River west of Godaleich Pond and in the walls of the tailrace below the Godaleich Pond powerhouse.

Lithology

The typical lithology of the Salmon River Dam Formation in the Great Burnt Lake area is fine grained, dense, thickly bedded sandstone. Fresh surfaces are typically grey to black, and locally have a purplish tinge. The surfaces commonly show fine laminations. Beds up to 30 cm thick, locally graded and containing cross-laminations, ripples and scour fill are characteristic. Outcrops commonly contain little or no pelitic material.

North of Godaleich Pond, the contact between the Salmon River Dam Formation and subunit 9a of the North Steady Pond Formation to the north (see below) is exposed in the tailrace below the Godaleich powerhouse. This contact is apparently conformable, gradational over approximately 100 m, and somewhat modified by later shearing. Facing directions in this exposure are ambiguous, complicated by local reversals due to short-wavelength isoclinal folds and overprinting by later shearing. However, unambiguous top determinations based on various sedimentary structures were possible on many outcrops on both the Salmon River Dam Formation and in the North Steady Pond Formation to the north and more than 80 percent of these are north-facing. This suggests that the overall succession faces to the north and that

the North Steady Pond Formation overlies the Salmon River Dam in this area. This is in accord with conclusions reached by Colman-Sadd (1976, 1980b) based on regional stratigraphic considerations in the St. Alban's (1M/13) and Twillick Brook (2D/4) map areas.

In thin section, the Salmon River Dam Formation is seen to consist mainly of quartz, feldspar, biotite and sericite. Polygonized quartz and subordinate amounts of feldspar form a very fine grained mosaic, which comprises up to 60 percent of the rock. Quartz also occurs as subrounded to rounded detrital grains up to 0.2 mm in diameter. These contain subgrain boundaries, are variably concentrated in sedimentary laminae, and commonly form up to 20 percent of the rock. Brown, pleochroic biotite is the next most common mineral. It forms equant grains and laths from 0.05 to 0.3 mm across, which are also variably concentrated in sedimentary laminae that comprise generally from 10 to 35 percent of the rock. The grain margins are sparsely sericitized and the grains show a weak but definite preferred orientation which is not always parallel to sedimentary stratification. Sericite and minor chlorite are ubiquitous but minor components, forming up to 15 percent in some samples. Sericite is more common, consisting of weakly oriented laths throughout the matrix; together, sericite, chlorite and biotite orientation define the principal cleavage direction (see below). Accessory minerals are zircon, epidote and/or clinozoisite and opaques.

Very fine sedimentary laminae of the order of 1 to 2 mm thick are well preserved in thin section (Plate 2-7). These are commonly defined by variations in the quartz+feldspar to ferromagnesian+opaque ratio. Some laminae consist of 90 percent very fine grained quartz with very minor ferromagnesian minerals. Where laminae have sharp boundaries, biotite and sericite are seen to overgrow them.

Contact Relations

The northern (upper) contact, exposed in the Godaleich tailrace, is interpreted as conformable, although modified by later reverse faulting and folding. The lower contact is not exposed in the present map area and has not been observed in adjacent map areas (Colman-Sadd, 1976, 1985a).

Structure and Metamorphism

The only penetrative fabric noted in outcrops of the Salmon River Dam Formation is an east- to east-northeast-trending cleavage which dips to the north at 55° to 70°. This cleavage is related to tight isoclinal folds; good examples of these folds are exposed in outcrops along the West Salmon River near its outflow into Round Pond. In this outcrop, the folds have an amplitude of 10 to 40 cm and are commonly sheared off on the limbs.

No major faults were mapped within the sequence although there are numerous small reverse faults exposed in the Godaleich tailrace.



Plate 2-7: Photomicrograph of sedimentary laminae in Salmon River Dam Formation sandstones. Biotite-rich laminae (coarse grained) alternate with fine grained quartzofeldspathic layers. Field of view is 6 mm.

North Steady Pond Formation (Unit 8)

Definition, Subdivision and Distribution

The North Steady Pond Formation was defined by Colman-Sadd (1980a) in the Twillick Brook (2D/4) map area where it consists principally of volcanoclastic rocks, lesser amounts of felsic volcanic rocks and minor mafic volcanic rocks. It has since been traced into the Burnt Hill (Colman-Sadd, 1985a) and Miguels Lake (Colman-Sadd and Russell, 1982) map areas.

Several lithologies have been recognized within the North Steady Pond Formation in the present area (see below), including a prominent felsic volcanic flow and pyroclastic sequence which is correlated with a similar felsic volcanic member, mapped by Colman-Sadd (1985a) in the Burnt Hill area to the east.

PIPESTONE POND AREA

Rocks assigned to the North Steady Pond Formation outcrop in a wide, east to west-trending belt across the southeastern part of the area. In the west, they are continuous with lithologies mapped by Colman-Sadd (1980a) in the adjacent Twillick Brook (2D/4) map area. Along this outcrop belt, the formation is bounded to the south by sandstones of the Salmon River Dam Formation and to the north by the western boundary fault of the Pipestone Pond Formation. In the area south of Sitdown Pond, the geological and structural grain of the rock units changes gradually to a more northwesterly and finally northerly direction, following the trend of the Pipestone Pond Complex boundary fault. In this area, the southwestern and western boundary of the formation is a fault along which it is juxtaposed with the newly defined Cold Spring Pond Formation (see below). The northern limit of the North Steady Pond Formation in this area is marked by a complex series of faults, which juxtaposes it with the Pipestone Pond Complex ophiolitic rocks.

Four lithological subunits have been recognized within the North Steady Pond Formation: subunit 8a—green, fine grained, siliciclastic sedimentary rocks; subunit 8b—massive, fine- to medium-grained, siliciclastic rocks; subunit 8c—polymictic, pebble to boulder conglomerate; subunit 8d—felsic pyroclastic and flow rocks. These subunits are the same as units 10, 11, 12 and 14 referred to by Swinden (1981b). The subunits described here are lithological subunits only; no stratigraphic sequence is implied by their order of listing and separate outcrop areas of the same lithological subunit are not necessarily stratigraphically equivalent. Each subunit will be individually described in the following section.

Subunit 8a

Lithology and Distribution. Subunit 8a consists mainly of well-bedded, graded sandstone and siltstone. Typical specimens are generally dark green, less commonly grey or brown, and exhibit fine laminations and partial Bouma sequences (Bouma, 1962). Subunit 8a lithologies along the southern edge of the North Steady Pond Formation outcrop in a 700 to 1000 m wide, east to west-trending belt north of Round and Godaleich ponds. In this area, well-developed partial Bouma sequences are present, consisting of graded psammite beds up to 4 cm thick overlain by thin, laminated, silty and slaty intervals, generally from 1 to 3 cm thick. These are best exposed in the stream bed above the Godaleich powerhouse and on hillocks north of Godaleich and Round ponds, and appear to be identical to North Steady Pond Formation turbidites described by Colman-Sadd (1980a) in the Twillick Brook (2D/4) map area to the east. They conformably overlie Salmon River Dam Formation sandstones to the south and appear to pass conformably and gradationally up into arkoses of subunit 8b to the north.

Similar lithologies also form a prominent band south of Sitdown Pond along the North Salmon River. In this area, typical lithologies consist of graded sandstone beds, interbedded siltstone and pebbly conglomerate and, in the upper parts of the subunit, shale. The lowermost part of this

sequence occurs at its northern contact where graded sandstones, pebbly conglomerates and lesser amounts of siltstone pass gradationally down into coarse polymictic conglomerates of subunit 8b. These beds commonly display very well-developed Bouma sequences, consisting of a basal graded pebbly sandstone approximately 60 cm thick succeeded by 50 to 70 cm of massive sandstone, and ultimately by 8 to 10 cm of laminated, and in places, cross-laminated siltstone and shale. The fine grained tops of these sequences are commonly scoured and rippled at the interface with the base of the overlying sequence. Locally, rip-up clasts and flame structures are seen. The succession tends to fine upward (i.e., northward); the stratigraphically highest exposures are those along the North Salmon River south of the junction with the brook draining Sitdown Pond; these are principally siltstone and shale rhythmically interbedded with 2 to 5 cm thick sandy beds. The northern gradational contact with the underlying conglomerate strikes into and is truncated by the west boundary fault of the Pipestone Pond Complex near the North Salmon River. The southern contact with felsic volcanic rocks of subunit 8d is a fault, which is exposed on the shore of the North Salmon River as a zone of intense shearing.

Lenses of subunit 8a lithologies, which are found within subunit 8b north of Godaleich Pond, are poorly exposed, and their relationship with the surrounding rocks is not entirely clear. They may represent stratigraphic lenses of shaly turbidite within the sequence. Alternatively, they may represent underlying or overlying rocks exposed in the hinge areas of folds or in fault bounded slivers. There is insufficient field evidence to evaluate these alternatives.

A narrow band of subunit 8a lithologies is present approximately 1.5 km northeast of Pipestone Pond. These rocks are poorly exposed but seem to consist mainly of fine grained chloritic schist interbedded with pebbly arkose.

In thin section, sandy lithologies of subunit 8a comprise a relatively coarse grained mosaic of polygonized quartz and feldspar grains. Relative proportions of the two minerals vary although quartz is generally dominant. The groundmass (matrix) forms from 40 to 65 percent of the rock. Chlorite is the principal ferromagnesian mineral, occurring as small laths and coarser (up to 0.2 mm) platelets surrounding the quartz and feldspar grains. Chlorite grains exhibit a weak to moderate orientation which defines the principal foliation. Texturally finer members of this subunit show a concomitant increase in chlorite content and generally stronger preferred orientations. Accessory minerals are rare, comprising mainly epidote, zircon and opaques, which commonly make up less than 0.5 percent of the rock.

Contact Relations. Contact relations of the two principal subunit 8a sequences between the West Salmon and North Salmon rivers are well documented. The exposed contact with the apparently underlying Salmon River Dam Formation in the Godaleich tailrace has already been described. The northern contact of the sequence north of Round Pond grades conformably into arkoses of subunit 8b, which are interpreted to overlie the turbidites in this area.

Turbidites of subunit 8a that outcrop along the North Salmon River south of Sitdown Pond conformably and gradationally overlie conglomerates of subunit 9c to the north. Beds of increasingly finer grained conglomerate persist for some distance southward into the turbidite sequence, and numerous top determinations in the turbidites uniformly indicate a south facing succession. This contact can be traced eastwards across the North Salmon River where it eventually strikes into the western boundary fault of the Pipestone Pond Complex. East of this point, subunit 8a is structurally juxtaposed with ophiolitic rocks and farther east with quartzose sedimentary rocks of the Spruce Brook Formation (Unit 12). The southern boundary of the turbidites appears to be everywhere faulted against felsic volcanic rocks of subunit 8d. This fault is well exposed on the shore of the North Salmon River as an intense shear zone several tens of metres wide; the shear zone (and presumably the fault) strikes slightly south of east in this area with variable dips within 5° of vertical. To the east, this fault eventually strikes into the Pipestone Pond Complex boundary fault. West of the North Salmon River, it swings to the northwest, cutting progressively lower stratigraphic levels of the turbidites until these rocks are pinched out between the fault and the underlying conglomerate.

The small band of subunit 8a northwest of Pipestone Pond is too poorly exposed for its contact relations to be certain. It appears to be faulted against mafic volcanics and pyroxenite of the Pipestone Pond Complex to the southeast. Its northwestern contact is in part faulted against gabbro, although part of this contact may be conformable with tuff and arkose of subunit 8b to the west.

Structure. Subunit 8a in all areas contains a weak to strong penetrative cleavage, defined in thin section by orientation of platy minerals (see above). In exposures between the West Salmon and North Salmon rivers, this cleavage trends approximately east-west, dips steeply to the north and is related to tight, upright F_1 isoclinal folds. In the exposures northwest of Pipestone Pond, this foliation is northeast-trending and dips steeply to the northwest. It is presumed to be related to F_1 isoclinal folds although none were observed in this area.

Subunit 8b

Lithology and Distribution. Subunit 8b is the most widespread lithological type in the North Steady Pond Formation. It outcrops in a wide, east to west-trending belt north of Round and Godaleich ponds, and again in a series of fault-bounded blocks southwest and west of Pipestone Pond. The principal lithologies in subunit 8b are epiclastic, arkosic sandstones and conglomerates, which are commonly very resistant and form prominent outcrop hills. The matrix of these rocks consists of aphanitic, medium- to pale-green, very siliceous material. Clasts visible in hand specimen comprise 10 to 50 percent of the rock and consist of feldspar and quartz crystals and crystal fragments with feldspar

commonly being the most abundant. Locally, up to 5 percent lithic clasts are present. These are predominantly white to pale green felsic volcanics, but minor amounts of pelitic sedimentary rock fragments also occur. Crystals, crystal fragments and lithic clasts visible in hand specimen generally range in size from 0.5 to 2 mm.

In some areas, the arkoses are interbedded with coarser grained rocks, ranging from pebble to cobble conglomerates. The matrix of these conglomerates, and also those assigned to subunit 8c, are identical to many subunit 8b lithologies. It appears that there is a continuum in composition and grain-size range from subunit 8b arkose to subunit 8c conglomerate.

Primary structures of any kind are rare. Bedding is rarely observed; where present, it takes the form of a faint banding, which reflects subtle differences in clast size and composition. Most outcrops of subunit 8b are massive and featureless.

Thin section examination of the rocks confirm most of the compositional features noted in hand specimen. Most specimens consist principally of 25 to 45 percent large (0.5–3 mm), subangular to subrounded quartz and feldspar grains in a siliceous, chloritic and sericitic matrix (Plate 2-8). The quartz grains are clear and usually without subgrain boundaries. The feldspar grains are likewise commonly single crystals, more often than not untwinned, and strongly altered, giving a very dusty appearance in plane light. Most contain inclusions of apatite, actinolite and, rarely, biotite. Some large crystals exhibit Carlsbad twins; albite and pericline twins are more commonly found in angular, equant to rectangular grains of a smaller size range (less than 0.5 mm), which comprise up to 40 percent of the plagioclase population in some sections. Lithic fragments are rare in most sections; where present, these consist mainly of very fine grained polygonized quartz and feldspar with minor sericite and chlorite and appear to be of felsic volcanic origin. The matrix, normally comprising less than 10 percent of the rocks, consists of very fine grained quartz and feldspar with varying amounts of sericite, chlorite, epidote and actinolite in order of abundance. Quartz and feldspar occur as an aphanitic mosaic. Fine grained sericite and chlorite are present in varying proportions (generally sericite greater than chlorite) and may comprise from 10 to 60 percent of the matrix. Sericite flakes and chlorite plates are normally preferentially aligned and define the principal F_1 foliation. Chlorite also occurs as larger plates throughout the matrix. Epidote occurs as euhedral crystals scattered throughout the matrix and with quartz and sphene, and locally pseudomorphs of biotite. Opaque minerals, sphene, apatite and, locally, zircon are accessories.

Sedimentary structures are no more obvious in thin section than in outcrop; however, some general observations may shed light on the origin of these rocks. The larger grains (crystals, crystal fragments and lithic clasts) range in shape from subangular to subrounded, and the degree of rounding is generally uniform within a single thin section. The larger clasts exhibit a moderate sorting but are commonly randomly oriented in the rock. This is generally true whether the clasts are so abundant that they are in mutual contact (clast

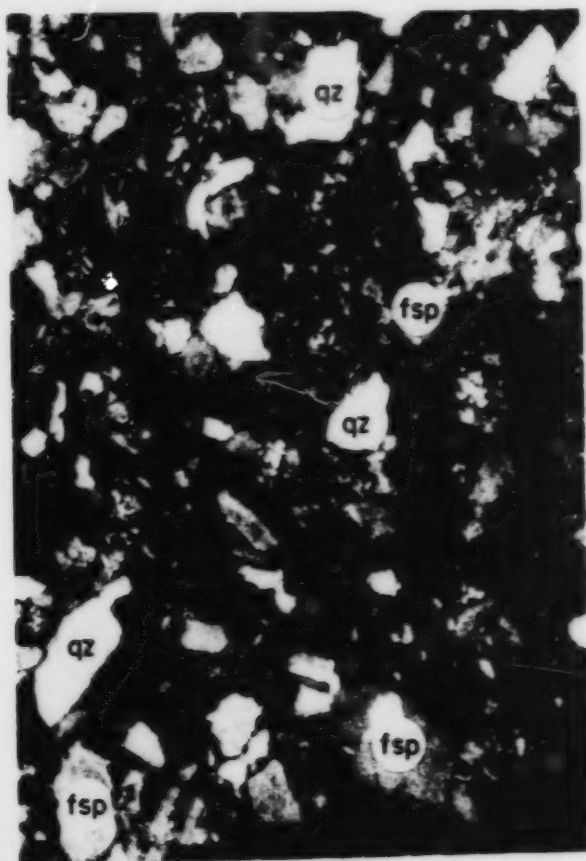


Plate 2-8: Photomicrograph of subunit 8b feldspathic (arkosic) wacke consists mainly of poorly sorted, subrounded to subangular quartz (qz) and feldspar (fsp) grains. Field of view is 6 mm.

supported) or whether they are isolated from one another in the matrix. The extreme angularity of (especially feldspar) grains in many specimens strongly suggests that little or no reworking by water has taken place. In addition, the lack of obviously water-laid material coupled with the generally volcanic appearance of the rocks in outcrop and thin section argues that most lithologies of subunit 8b are ultimately tuffaceous in origin. Some reworking may be indicated by crude stratification in some outcrops, local interbeds of volcanic derived conglomerates, the apparent compositional and grain size continuum between subunit 8b and conglomerates of subunit 8c and by the variable amount of rounding from one section to another. However, no reworking would appear to be demanded by the evidence and the above features, (particularly the lack of internal structure over considerable thicknesses of section), are better explained by attributing these deposits to mass flow processes on the slopes of volcanic edifices. These may be analogous to the massive sandstone facies described by Walker (1978).

Contact Relations. The band of subunit 8b north of Round and Godaleich ponds has conformable contacts to both the north and south. To the south, it is apparently gradational over a limited stratigraphic interval with turbidites of subunit 8a. It is presumed to overlie the turbidites, mainly on the basis of a predominance of northerly facing directions in subunit 8a, and this notion gains some support from a few north-facing directions in graded beds in subunit 8b. The northern contact is conformable with felsic volcanic rocks of subunit 8d; this contact is generally marked by a thin (a few metres) interval of fine grained, dark grey to black clastic sediments. To the west, these rocks strike into and are truncated by the eastern boundary fault of the Cold Spring Pond Formation.

Subunit 8b west of Pipestone Pond occurs in a series of structural blocks, the largest of which is bounded to the east and north by the western boundary fault of the ophiolite. Several smaller blocks are bounded by smaller faults, which are generally defined by slivers of ophiolitic rock outcropping along their trace. Most of these structural blocks are bounded to the west by the eastern boundary fault of the Cold Spring Pond Formation.

Structure. Most outcrops of subunit 8b are massive and not obviously deformed. However, close inspection commonly reveals a single penetrative fabric probably related to the F_1 isoclinal folds observed in subunit 8a. North of Round Pond, this cleavage strikes east to west but farther west in the area north of Godaleich Pond, it has a more northeasterly trend. In these areas, the foliation is defined by platy mineral alignment. Farther to the northwest (west of Pipestone Pond), subunit 8b contains a stronger penetrative foliation defined by both mineral alignment and by stretching of crystal and lithic grains in the plane of the foliation. No minor folds were observed in this subunit but the foliation parallels that in the adjacent turbidites of subunit 10a where isoclinal folds are commonly observed.

Subunit 8c

Lithology and Distribution. Subunit 8c consists of polymictic conglomerate and minor interbedded sandy (sometimes turbiditic) beds. It is of relatively limited aerial extent within the North Steady Pond Formation, and is most widely distributed in a prominent band along the North Salmon River south of Pipestone Pond. Similar rocks also occur interbedded with arkosic rocks of subunit 8b in several fault bounded blocks southwest and west of Pipestone Pond, and as lenses in the prominent belt of subunit 8b lithologies north of Godaleich Pond and the West Salmon River. Conglomerates in all areas are commonly matrix supported and contain from 25 to 75 percent clasts (Plate 2-9). The matrix is commonly pale to medium green and fine grained, although in some exposures it is pebbly, containing visible grains of quartz and feldspar (crystals and crystal fragments) from 0.5 to 3 mm in size. In the latter case, the matrix looks very much like some crystal-rich varieties of subunit 8b. Gravel-sized fragments range from less than 1 to 50 cm (locally in the same outcrop), although they most commonly fall in the 3 to 15 cm range. They represent a considerable

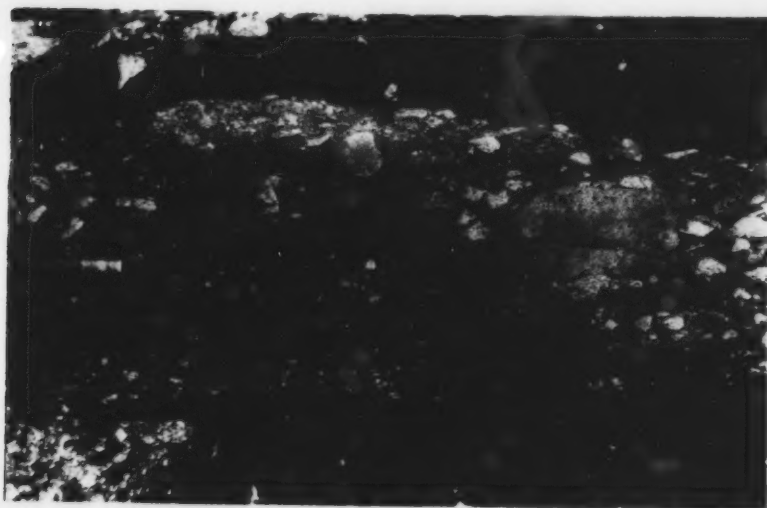


Plate 2-9: *Polymictic conglomerate of subunit 8c. Clasts include mafic and felsic volcanic rocks, arkose, tuff and minor gabbro. Matrix is similar to arkose depicted in Plate 2-8.*

variety of rock types; the most common lithologies represented are felsic tuff, generally white to buff weathering and containing prominent quartz and lesser amounts of feldspar phenocrysts, and pelitic sedimentary rocks. Mafic volcanic rocks are generally present and, although they form a relatively minor component, they are locally the dominant clast type (e.g., in outcrops along the North Salmon River adjacent to the contact with basalts of the Pipestone Pond Complex). Granitic cobbles, which are locally pegmatitic and commonly resemble plagiogranite, are present in some outcrops as are gabbro and diabase. Some, but not all granitic and gabbroic clasts are both larger and more rounded than the rest of the clast assemblage suggesting that these may in part be derived from a pre-existing sedimentary source. The clast assemblage does not appear to show any systematic trends between the various areas where it is exposed with the exception of the previously noted concentration of mafic volcanic clasts in outcrops along the North Salmon River.

Coarse fragments within the conglomerates are generally angular to subangular, randomly oriented within the rock and are very poorly sorted. Even the very tabulate clasts show little tendency to be preferentially oriented within the rock suggesting that they were probably not deposited by flowing water. Stratification is not normally developed on outcrop scale; however, some outcrops along the North Salmon River do exhibit a very crude stratification, marked by discontinuous layers ('beds') of boulder-rich, clast-supported material up to 30 cm thick or by a subtle grading of clasts across a 1 to 2 m interval. In a southeasterly direction along the river, the conglomerates are seen to assume a more orderly aspect near the gradational contact with the turbidites of subunit 8a. Interbeds of pebbly material appear and become progressively more common. Recognizable strata with abundant scour channels and rip-up clasts are developed and the average size of the clasts decreases.

In thin section, these rocks are seen to consist of a matrix of fine quartz and feldspar with fine platelets of chlorite and sericite, which form a relatively uniform mosaic. Occasionally, this mosaic also contains isolated larger grains of quartz and feldspar up to 1 mm, which are commonly subangular to subrounded crystal fragments. Epidote is a rare constituent forming euhedral grains in the matrix and occurring as an alteration mineral after plagioclase. Biotite and actinolite are also present as relatively minor alteration phases. In contrast to most other lithologies of the North Steady Pond Formation, veins and individual crystals of calcite form a prominent alteration phase. Clast assemblages viewed in thin section generally confirm the field observations. Quartz- and feldspar-porphyrific felsic volcanic and pyroclastic rocks predominate in most sections. Samples from along the North Salmon River close to the basalt contact tended to be richer in basalt clasts. Mineralogy and textures of the clasts do not appear to be significantly different from outcrops of the various lithologies and are therefore not described in more detail here.

The most striking character of these conglomerates, both in outcrop and in thin section, is their internal chaos. The wide variety of provenances for the clasts, the general lack of sedimentary structure, poor sorting, angularity of most clasts and the matrix support, all suggest very rapid deposition, probably as debris flows.

Contact Relations. Along the North Salmon River, contact relations of subunit 8c are clearly observable. Top determinations in this unit, usually from graded bedding, indicate a uniformly south- to southwest-facing sequence (see cross-section B-B₁; map in pocket). The southeastern contact with the overlying turbidites of subunit 8a is conformable and gradational as previously described. The northwestern extension of this contact strikes into and is truncated by a

major northwest-trending fault, which ultimately cuts the upper levels of the conglomerate succession, juxtaposing it with probably north-facing felsic volcanic rocks of subunit 8d. The nature of the northern contact with basalts of the Pipestone Pond Complex (Unit 6_p) is more enigmatic and contact was not observed in the field. However, several arguments suggest that the contact is not structural: 1) bedding in the conglomerates is generally subparallel to bedding in the basalt inferred from pillow orientations; 2) close to this contact, detritus in the conglomerate is predominately basalt and gabbro, suggesting derivation from the Pipestone Pond Complex; and 3) there is no evidence of shearing or development of a tectonic fabric parallel to the contact. Prominent shear zones are present but in all cases trend roughly northeast-southwest, i.e., at right angles to the contact. These shear zones appear to be related to transverse faults which postdate the conglomerate-basalt contact and offset it in several localities. They also parallel and are probably related to faults which disrupt the stratigraphy of subunits 8b and 8c southwest and west of Pipestone Pond; the present evidence, therefore, would seem to favour a sedimentary contact in this area such that conglomerates of subunit 8c overlie (probably disconformably) basalt of the Pipestone Pond Complex.

Southwest and west of Pipestone Pond, the conglomerates are relatively poorly exposed and their contact relations are less certain. They appear to be conformably interbedded with other lithologies of the North Steady Pond Formation, notably arkoses of subunit 8b and locally turbidites (subunit 8a).

North of Godaleich Pond and the West Salmon River, conglomerates form at least two relatively large lenses within the outcrop band of subunit 8b. These are likewise not well exposed. They may represent sedimentary lenses within the arkoses; alternatively, they may overlie or underlie subunit 9b and be exposed in the hinge areas of F_1 folds, the axes of which parallel the long direction of the lenses in this area.

Structure. Deformation in subunit 8c is most intense south and west of Pipestone Pond. In these areas, a single penetrative, northeast-trending foliation is variably developed, which dips steeply to the northwest; it is defined by alignment of sericite and chlorite platelets and locally by stretching of lithic clasts. In areas where bedding can be recognized, it appears to be subparallel to this foliation, and the fabric is therefore probably related to the F_1 isoclinal folds that affect other parts of the North Steady Pond Formation. An east-west trending, north-dipping foliation in lenses of conglomerate north of Godaleich Pond is probably also related to this deformational event.

Subunit 8d

Lithology and Distribution. Subunit 8d consists of predominantly felsic volcanic rocks, which underlie a series of hills south of the North Salmon River and extend eastwards

to the boundary of the map area. These rocks are identical to and continuous with similar rocks that have been described in the Twillick Brook (2D/4) and Burnt Hill (2D/5) map areas to the east (Colman-Sadd, 1980a; 1985a) respectively, and are informally referred to as the Huxter Pond Volcanic belt by Colman-Sadd and Swinden, (1982). These rocks are particularly well exposed in well-washed exposures along the shores of the North Salmon River.

Three principal lithologies have been recognized in this subunit in the Pipestone Pond area. In decreasing order of abundance these are (with descriptions):

1) Massive, aphanitic, dense, very siliceous rhyolite: this rock ranges in colour from buff and pale green to dark grey and black. Flow banding and flattened pumice fragments are seen in some exposures; spherulites are well developed in at least one locality. Locally, this unit is so siliceous that it has a cherty aspect.

2) Feldspar-quartz porphyry: this rock is less silicic-appearing than the aphanitic varieties, generally being buff to pale or olive green in hand specimen, and containing 1 to 15 percent white feldspar crystals up to 2 mm and fewer (and generally smaller) quartz phenocrysts. In some localities, it is very amygdaloidal, the amygdales being filled principally by chlorite, giving it a distinctive spotted appearance. This rock has been termed a dacite porphyry as a field term and occurs principally in the northeastern parts of the unit. It seems likely that more detailed mapping could establish it as a mappable stratigraphic horizon. Flow banding is recognized locally as are autobrecciation and devitrification textures.

3) Medium- to coarse-grained felsic pyroclastic rocks: these rocks consist of angular to subrounded fragments of porphyritic rhyolite in a fine grained matrix composed of silicic tuff (Plate 2-10). The breccias locally are quite coarse and contain chaotic assortments of fragments up to 40 cm in size randomly mixed together, perhaps suggesting the proximity of an explosive vent (or vents).

The felsic volcanics are locally interbedded with dark-grey to black argillites, and are separated from the tuff and arkose of subunit 8b to the southwest by a thin band of grey to black pelites, which are intensely deformed and too thin to be mapped at the present scale.

In thin section, the aphanitic rhyolites consist of a very siliceous groundmass of fine grained, granoblastic quartz and feldspar (Plate 2-11). Sericite and chlorite occur as fine flakes in amounts generally less than 10 percent. In one section, faint crystal outlines and a slight compositional variation were noted which appear to represent a completely altered feldspar phenocryst. In some sections, quartz and feldspar crystals up to 0.1 mm, locally embayed, are present in the groundmass (Plate 2-11); these may be either microphenocrysts or porphyroblasts formed during recrystallization due to metamorphism.

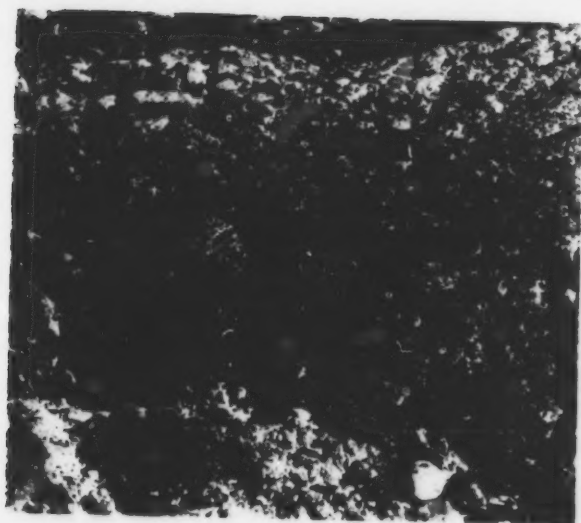


Plate 2-10: *Felsic volcanic breccia in subunit 8d consisting of large rhyolitic fragments in a rhyolitic tuffaceous groundmass. Pocket knife near left edge of photo for scale.*

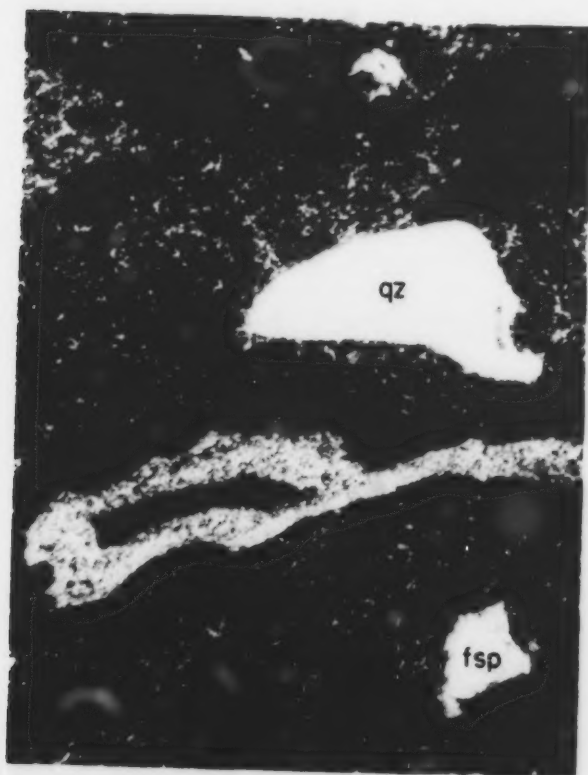


Plate 2-11: *Photomicrographs of Subunit 8d rhyolite shows rounded, embayed quartz phenocryst and a collapsed pumice fragment. Field of view is 6 mm.*

Porphyritic dacites show a more variable aspect in thin section. They generally consist of 65 to 85 percent fine grained groundmass with the remainder a phenocryst assemblage of quartz, feldspar and rarely, altered mafic minerals. The groundmass consists of an assemblage of recrystallized feldspar, commonly as very fine, randomly oriented laths, polygonized quartz and a mixture of varying proportions of sericite and chlorite. Relative abundance of chlorite and sericite in the groundmass is widely variable: sericite was very rarely seen in one section, whereas in others it is more abundant than chlorite by up to 1.5:1. The relative amount of chlorite+sericite with respect to quartz+felspar also varies considerably, ranging from 1.5:5 to 5:1; increasing proportions of the platy minerals correlate with progressively darker colour of the rock. The groundmass locally contains secondary growth of epidote and sphene. Zircon and apatite are accessory primary phases.

Phenocrysts in the dacite porphyry are predominantly feldspar with lesser amounts of quartz. Feldspar phenocrysts commonly show a bimodal size distribution. Small plagioclase laths up to 0.3 mm long are randomly oriented in the matrix of many samples, and may form up to 40 percent of the phenocryst feldspars present, although their abundance is usually less than 20 percent. Larger feldspar phenocrysts, up to 4 mm in size, are the more abundant phase; they comprise both orthoclase and plagioclase, are generally equant, rounded and embayed and commonly have reaction rims of micrographic quartz. These two feldspars probably represent two phases which are present during solidification of the melt, an equilibrium plagioclase phase represented by the small laths and a disequilibrium (cumulate?) feldspar phase. All feldspars are moderately to strongly saussuritized, characterized by the secondary development of sericite and quartz or, less commonly, chlorite-sphene-quartz. There is some evidence for the presence of a more mafic phenocryst phase; assemblages of epidote-sphene-quartz are locally present and appear to pseudomorph biotite.

Felsic volcanic clasts are locally present in the dacite porphyry. They are usually subrounded and consist almost wholly of quartz, feldspar and sericite. The largest clast observed in thin section was approximately 2 mm across.

Original volcanic textures are not well preserved in thin sections of these rocks. The most prominent textural feature observed is flattened vesicles which are abundant in some sections. These commonly exhibit a concentrically zoned infilling, consisting of aphanitic chlorite and variable epidote around the margins followed inwards by a zone rich in sericite and chlorite and, in some cases, a core of granoblastic quartz. One vesicle was filled only by granoblastic quartz. This zoning is consistent both within a single section and in sections from separate areas indicating a fairly homogeneous evolution of late-stage fluid phases from which the vesicle filling precipitated. The vesicles range in size from 1 to 5 mm and are generally slightly to moderately flattened.

One thin section of epiclastic rock from this subunit was examined. It comprises a relatively fine grained assemblage of moderately sorted, subrounded to subangular quartz and feldspar clasts in a siliceous, fine grained matrix. The clasts are dominantly crystals and crystal fragments (feldspar:quartz = 60:40) ranging in size from 0.5 to 3 mm. A few rhyolitic lithic clasts were also noted. The matrix makes up about 30 percent of the rock and consists of chlorite, quartz, feldspar and minor sphene and epidote.

Field observations of features such as flow banding, autobrecciation, and flattened pumice fragments, coupled with thin section observations of partially resorbed phenocrysts and the local presence of lithic clasts, suggest that most of the porphyritic dacite, and perhaps some of the aphanitic rhyolite formed as ash flows during an episode of subaerial volcanism. This interpretation is strengthened by the lateral extent of this unit; it can be traced for over 30 km to the east and northeast of the present area (Colman-Sadd, 1980a, 1985a). At least some of the aphanitic rhyolites do not exhibit any pyroclastic textures in outcrop or thin section and it is possible that they formed as rhyolitic flows. The presence of coarse pyroclastic rocks suggests that local centres of explosive volcanism existed during production of this unit and the intercalation of apparently marine sedimentary rocks suggests that much of this volcanism was not far removed from sea level.

Contact Relations. To the north, the felsic volcanic rocks are faulted against south-facing turbidites and conglomerates of subunits 8a and 8c along a major, east to southeast trending structure, which outcrops in the North Salmon River. To the south, they are apparently in conformable contact with arkoses of subunit 8b, separated from them by a thin band of dark grey to black, pelitic sediment. In accordance with arguments developed during the discussion of structure in subunits 8a and 8b north of Godaleich and Round Ponds, the felsic volcanics are interpreted to overlie subunit 8b in this area, and to cap an almost continuous north-facing sequence, which is terminated against the fault that marks the north boundary of subunit 8d (see cross-section B-B₁; map in pocket). The conformable contact between the felsic volcanics and subunit 9b strikes into and is truncated by the eastern boundary fault of the Cold Spring Pond Formation. North of this point, the subunit 8d is juxtaposed with turbidites of the Cold Spring Pond Formation.

Structure. Structural features are not well developed in the massive, homogeneous, very siliceous rocks of subunit 8d. A fracture cleavage is locally developed, which is steeply dipping and trends east to west in the eastern part of the area and northeast to southwest in the area south of Pipestone Pond, conforming to better-developed trends in adjacent sedimentary rock units. In subunit 8d, the cleavage is generally related to local shearing, but in a regional sense, it is probably related to the deformation that produced F₁ isoclinal folds in the adjacent units. Weak to moderate alignment of sericite and chlorite in the groundmass of some rocks, and slight to moderate flattening of vesicles probably further reflect this deformational episode.

Metamorphism

Secondary mineral assemblages of chlorite-*albite*-epidote in fine grained sedimentary rocks show that metamorphism of the Baie d'Espoir Group in this area took place under lower- to middle-greenschist-facies conditions. Alignment of platy minerals in the main foliation suggests that this metamorphism was earlier than, or synchronous with, development of the isoclinal folds.

Correlation and Age

There is no direct evidence in the Pipestone Pond area for the age of the Baie d'Espoir Group. Colman-Sadd (1976) reported fossils from the boundary between the Riches Island and the St. Joseph's Cove formations near Conne River, which indicate a Middle Ordovician (Late Arenig-Early Caradoc) age. Dunlop (1954) reported some poorly preserved fossils from the Riches Island and Isle Galet formations, which suggested an Ordovician or Cambrian age. Rocks of the Davidville Group near Middle Ridge, which are along strike with the Baie d'Espoir Group, contain Middle Ordovician graptolites (Anderson and Williams, 1970). Stratigraphic continuity can be demonstrated between the North Steady Pond and Salmon River Dam formations in the present area and the main body of the Baie d'Espoir Group to the east (e.g., see Colman-Sadd and Swinden, 1982, their Figure 3); there seems to be little doubt that the Early to Middle Ordovician age of the Baie d'Espoir Group is applicable to these formations in the present area.

UNIT 9

Definition and Distribution

Unit 9 consists of metasedimentary and metavolcanic (?) rocks that underlie an area northwest of Gulp Pond. The rocks form a northeasterly trending belt approximately 2 km wide, which extends from the area west of South Pond northward to the northern edge of the map area. Similar lithologies have been mapped in the adjoining Noel Paul's Brook (12A/9) map area to the north (Kean and Jayasinghe, 1980).

Lithology

Two distinct lithological subunits of approximately equal apparent thickness are present within Unit 9, a polymictic, pebble to cobble conglomerate (subunit 9a) to the east and interbedded chloritic schist and possible mafic metavolcanic rocks (subunit 9b) to the west.

Subunit 9a

Subunit 9a consists of coarse, polymictic conglomerate. It is commonly poorly sorted and clast supported, containing 40 to 75 percent subrounded to subangular, 5 to 15 cm clasts in a grey, fine grained matrix. The clasts are principally massive to quartz-porphyritic rhyolites. Mafic volcanic rocks locally form up to 20 percent of the clasts and minor

sedimentary and felsic plutonic clasts are also present. As in the conglomerates of the North Steady Pond Formation to the south, granitoid boulders, where present, tend to be both larger and more rounded than the bulk of the clasts. Although these conglomerates structurally overlie the Pipestone Pond Complex, no mafic plutonic or ultramafic clasts seem to be present. Subrounded quartz grains up to 2 mm are locally abundant in the matrix, and at one locality north of Gulp Pond, a large (15 cm) red jasper clast was observed. Sedimentary structures are generally not developed, although crude bedding is locally defined by variations in grain size. The conglomerate is locally interbedded with quartzofeldspathic volcanoclastic rocks consisting of fine grained, granoblastic quartz and minor sericite, in which rounded quartz grains (approximately 5 percent) and some lithic clasts of both felsic and mafic volcanic rocks are set.

Subunit 9b

Subunit 9b is very poorly exposed, represented only by a few outcrops and boulder patches. The principal lithology seems to be a very fine grained green, chloritic, ordinarily fissile rock, which is commonly highly sheared. The rock may represent either a mafic tuff, volcanoclastic sediment or, in some cases, mafic flow. Original textures are not commonly preserved, although in one locality possible vesicles were noted, suggesting, at least in this outcrop, a volcanic origin.

Contact Relations

Unit 9 is fault bounded on all sides within the present map area. To the east, conglomerates of subunit 9a are juxtaposed with gabbro and ultramafic rocks of the Pipestone Pond Complex along a fault, which has a marked topographic expression west of Gulp Pond. To the west, mafic volcanic and sedimentary rocks of subunit 9b are bounded by a fault, which is marked by a large sliver of serpentinized peridotite northwest of Gulp Pond and another of (presumably ophiolitic) gabbro 2 km to the southwest. They are presumably faulted against foliated, porphyritic, granitoid rocks of Unit 13 in this area, although outcrop is very scarce and the bedrock lithology is uncertain northwest of the fault. To the southwest, Unit 9 appears to be pinched out between this fault and the western boundary fault of the Pipestone Pond Complex.

Structure and Metamorphism

Rocks of Unit 9 carry a strong penetrative cleavage, S_1 , which trends northeast to north-northeast and has a steep dip, variably to the southeast or northwest. This fabric is defined in the conglomerates by flattening of the clasts in the plane of the foliation and by the orientation of chlorite and sericite in the matrix. Locally, intense shear zones parallel this fabric; however, it is normally oblique to the major faults which bound the rock units. Bedding is seldom present but where seen, it is subparallel to the S_1 fabric. S_1 in this area parallels the principal foliation in similar rocks, both to the southwest in the present map area and also to the northeast in the

Miguels Lake area (Colman-Sadd and Russell, 1982). The fabric in both areas is related to the formation of upright isoclinal folds. These regional considerations, coupled with the apparently subparallel bedding-cleavage relationships in Unit 9, suggest that this is the same deformational episode.

Correlation and Age

There is no direct evidence within the Great Burnt Lake map area regarding the age of Unit 9. It appears to be stratigraphically continuous to the north with rocks in the Noel Paul's Brook (12A/9) map area, which Kean and Jayasinghe (1980, their Unit 3) infer on the basis of regional stratigraphic considerations to be Ordovician. They suggested that these rocks were time equivalents of the Davidsville and Baie d'Espoir groups. Probable stratigraphic equivalents in the Miguels Lake area have been referred by Colman-Sadd and Russell (1982, their Unit 11) to either the Baie d'Espoir or Victoria Lake groups, both of which are Early to Middle Ordovician based on fossil evidence (this report; see discussions of Baie d'Espoir Group and Cold Spring Pond Formation ages). In the Pipestone Pond area, lithological similarities can be seen between Unit 9 and both subunit 8c of the North Steady Pond Formation (Baie d'Espoir Group) and subunit 10b of the Cold Spring Pond Formation. They occupy a similar structural position as well, immediately 'outboard' of the Pipestone Pond Complex. These rocks are therefore considered to be correlative with either the Baie d'Espoir Group or the Cold Spring Pond Formation. Further definition of this age awaits more detailed correlations.

COLD SPRING POND FORMATION (UNIT 10)

Definition and Distribution

The name Cold Spring Pond Formation is formally proposed for sedimentary and volcanic rocks which underlie most of the southwest quadrant of the map area and extend as a narrow band northwards to the area immediately southwest of South Pond. These rocks have recently been traced westward beyond the limits of the present map area by Colman-Sadd (1984), (see also Colman-Sadd and Swinden, 1983). The name proposed here is taken from Cold Spring Pond, a large body of water which forms part of the West Salmon River system south of Great Burnt Lake. The rocks herein assigned to the Cold Spring Pond Formation have previously been informally referred to as part of the 'Great Burnt Lake volcanic belt' (Colman-Sadd and Swinden, 1982). However, this name is unsuitable as a formal designation because the formation as presently defined does not outcrop on Great Burnt Lake. The Cold Spring Pond Formation is not included in the Baie d'Espoir Group because the volcanic lithologies, comprise dominantly black silicic volcanics and extensive pillow lavas. Volcanic members of the Baie d'Espoir Group, in contrast, are characterized by white to buff felsic volcanic rocks and a general paucity of mafic volcanic lithologies (Colman-Sadd, 1980b; Colman-Sadd and Swinden, 1982).

PIPESTONE POND AREA

The Cold Spring Pond Formation comprises predominantly siliciclastic sedimentary rocks; the principal lithologies are green volcanoclastic arkose and greywacke, which commonly contain partial Bouma turbiditic sequences. Substantial volumes of black, graphitic slate and lesser amounts of polymictic conglomerate are interbedded with the turbidites. The formation is characterized, however, by a volcanic assemblage (up to 30 percent) consisting of laterally extensive mafic pillow lava and massive basalt associated with dense, black, very siliceous quartz ($\pm$ feldspar)-crystal tuff and rhyolitic porphyry. Sedimentary members of the Cold Spring Pond Formation are generally similar to those of the North Steady Pond Formation, but the two formations are distinguishable on the basis of their distinctly different volcanic rock assemblages, and by the absence in the Cold Spring Pond Formation of massive, unbedded arkose similar to subunit 8b. The Cold Spring Pond Formation is distinguished from the Salmon River Dam Formation by the presence of volcanic rocks, volcanic detritus and partial Bouma sequences.

The Cold Spring Pond Formation is fault bounded throughout its presently mapped extent. The eastern margin is marked by a major north- to northeast-trending fault, which juxtaposes it with the Salmon River Dam Formation in the southern corner of the map area, with the North Steady Pond Formation throughout the southern and central parts of the map area and with the Pipestone Pond Complex west of the north end of Pipestone Pond. The western margin of the Cold Spring Pond Formation is likewise marked by a fault, which juxtaposes it with schist, gneiss and granite of the western igneous-metamorphic domain, and, in the south, with ophiolitic rocks.

Good representative exposures of all major rock units in this formation can be viewed in road cuts and quarries along the North Salmon Dam access road between Godaleich Pond and the North Salmon Dam, so that this is designated as a reference section for the formation. Typical sections of the various lithological subunits will be identified in the following descriptions of the individual subunits.

Four lithological subunits have been recognized within the Cold Spring Pond Formation: subunit 10a—green volcanoclastic rocks and interbedded black graphitic shale; subunit 10b—conglomerate; subunit 10c—quartz-porphyrific, felsic volcanic rock; subunit 10d—basalt and basaltic tuff. These subunits are described separately below. No stratigraphic order is implied by the order in which they are presented and separate outcrop areas of the same lithology are not necessarily stratigraphically equivalent.

No meaningful estimate of the true thickness can be offered due to the tight folding of strata and scattered outcrop.

Subunit 10a

Lithology and Distribution. Subunit 10a consists of fine- to medium-grained siliciclastic sedimentary rocks. These form the greatest volume of rocks within the Cold Spring Pond

Formation, and are most widespread in a broad belt along the eastern side of the formation's outcrop area. These rocks have been traced from the area of the North Salmon River east of Great Burnt Lake south and southwest to the limits of present mapping near Cold Spring Pond and the West Salmon River. Excellent exposures of the sedimentary rocks, which characterize this part of subunit 10a, can be seen in the bed of the West Salmon River east of Cold Spring Pond and in road cuts and quarries along the North Salmon Dam access road. Sedimentary and tuffaceous rocks assigned to subunit 10a also outcrop in two narrow bands north of the North Salmon River.

The principal lithologies in subunit 10a south of the North Salmon River are green, epiclastic pelite and psammite interbedded with lesser amounts of carbonaceous pelite. The epiclastic rocks are commonly well bedded; average bedding thickness is of the order of 10 cm in most areas but is usually widely variable, ranging up to 50 cm. Psammite beds are locally graded and contain scour channels at the base; slump (soft sediment) folding, flame structures and rip-up clasts are locally seen on good exposures. Delicate structures are not generally visible on weathered outcrops but unweathered surfaces in some quarries reveal fine laminations at a scale of less than 1 mm in the more siliceous members. Locally, quartz and feldspar grains up to 1 mm can be seen in hand specimen. Near Cold Spring Pond average grain size is generally coarser, although, paradoxically, this appears to be accompanied by an increase in the thickness and frequency of pelite interbeds as well. Graded, pebbly and conglomeratic beds are common in this area and are generally rhythmically interbedded with green pelitic intervals 5 to 10 mm thick. The pelitic layers are commonly too fissile to show any original textures in outcrop.

Throughout the sequence black graphitic pelite is a ubiquitous component. It is generally interbedded with the green, epiclastic pelite and locally contains 1 to 2 percent pyrite. This lithology is very recessive and is not often exposed; drill logs of holes in the Cold Spring Pond and Prawn Pond areas report more of this lithology than is normal for outcrop sections, suggesting that black pelite may be more abundant in the section than suggested by its surface distribution.

Northwest of Pipestone Pond, the two narrow bands of subunit 10a consist of fine grained, dark-green, chloritic schist, which was probably originally volcanoclastic pelite and tuff derived from mafic volcanics. These rocks are, in general, considerably less siliceous than those to the south. They also contain interbedded black graphitic pelite.

Rocks of subunit 10a are seen in this section to be dominantly quartzofeldspathic. Very fine grained varieties consist of a mosaic of equant subrounded quartz and feldspar grains, generally less than 0.05 mm across, which are surrounded by slightly coarser grained green chlorite and, less commonly, sericite flakes. The chlorite grains range from medium to bright green and are strongly pleochroic. Sericite, when present, occurs as fine flakes. Thin laminae in the rock

are defined by variations in the chlorite:quartz+feldspar ratio and are of the order of 0.4 mm thick. In some sections, laminae are defined by lines of single chlorite flakes, arranged end to end in the plane of bedding, extending across the width of the slide and, locally, defining subtle cross-lamination as well. The proportions of the various minerals varies considerably from sample to sample and even within the same specimen. In general, however, quartz+feldspar is more abundant than chlorite+sericite, and chlorite is more abundant than sericite. In the area northwest of Pipestone Pond, where deformation is generally more intense, mafic minerals predominate over quartz+feldspar and biotite becomes a prominent phase, perhaps suggesting a slightly higher grade of metamorphism.

Increasing grain size in the psammitic beds in the area east and southeast of Great Burnt Lake is commonly reflected by increasing amounts of 1 to 3 mm-sized clasts, generally consisting of quartz (30 to 60 percent), feldspar (up to 30 percent) and lithic fragments (less than 10 percent). These grains tend to be subrounded (although some retain sharp corners) and fairly well sorted within individual beds.

Coarse sandstones in the Cold Spring Pond area contain framework clasts up to 2 mm in size. These form 70 to 80 percent of the rock. Clasts are subrounded to subangular and packed sufficiently tightly so that the rock tends to be clast supported. The matrix consists of fine grained, polygonized quartz and coarser (0.5 to 1 mm) chlorite flakes. Lithic clasts range up to 20 percent, comprising in increasing order of abundance, felsic volcanic clasts (quartz-feldspar-sericite-chlorite), chert or silicic tuff (granoblastic quartz) and, rarely, basalt.

Alteration is well advanced in most specimens where individual clasts can be resolved. Feldspars are generally very dusty, containing clear reaction rims and inclusions of chlorite, sericite and, less commonly, actinolite. In some sections, lithic clasts are altered to quartz-chlorite-sericite and quartz-chlorite actinolite assemblages. Calcite veining was noted in two sections. Accessory zircon, apatite and opaque minerals are generally present.

Subunit 10a appears to represent turbidites of volcanic provenance. No ophiolitic rocks were recognized in the pebbles of the Cold Spring Pond Formation sediments.

Contact Relations. Subunit 10a forms both the eastern and western boundary of the Cold Spring Pond Formation through much of its mapped extent. It is present along almost all of the eastern boundary, where it is faulted against sedimentary and volcanoclastic rocks of the North Steady Pond Formation and, in the extreme north, where it is faulted against ophiolitic rocks of the Pipestone Pond Complex. Along the western boundary, it is faulted against deformed granitoid rocks of Unit 13 in the area east of the north end of Great Burnt Lake. In the area north of Cold Spring Pond it is faulted against a sliver of ophiolitic rocks. Shearing along the boundary faults is locally intense and is marked in thin section by the development of spaced cleavage, wherein an

early planar fabric is isoclinally folded and the limbs strongly sheared out in the plane of the second deformation.

Within the Cold Spring Pond Formation, subunit 10a is apparently in conformable contact with all other subunits, underlying mafic volcanic rocks east and northeast of Great Burnt Lake and felsic volcanic rocks east and southeast of Great Burnt Lake, and overlying conglomerates along the shores of Cold Spring Pond.

Structure. Rocks of subunit 10a carry evidence of at least two deformations. In all areas, the principal fabric is a steeply dipping cleavage which is axial planar to upright, isoclinal folds. This cleavage trends approximately northeast in the northern and central parts of the area and easterly in the area near Cold Spring Pond. F_1 folds are rarely seen in the field but where present they commonly have an amplitude of 0.5 to 1 m and are sheared out along the limbs. This fabric is represented in thin section by alignment of chlorite ($\pm$ sericite) in the plane of the cleavage. A second deformation of variable intensity is locally recognized. In the area north of the Salmon River and west of Pipestone Pond, the first cleavage is folded around moderately open, asymmetrical folds with steep eastern limbs, the axes of which are approximately parallel to those of the F_1 folds. In the southern and southeastern parts of the area, however, the second deformation is represented by a steep crenulation cleavage which trends north to northwest and is mainly developed in fine grained sedimentary rocks. No folds associated with this cleavage were recognized.

Internal faults are not easily recognized within subunit 10a because of the relative uniformity of rock types within the sequence and the poor exposure. Evidence for the presence of such faults is the presence of several isolated slivers of ophiolitic rocks, ranging from ultramafic to gabbroic, which interrupt the sequence between the North Salmon River and Cold Spring Pond. The location of these faults and the relative merits of interpreting them as steeply dipping sets of thrust faults versus relatively flat thrusts have been discussed in a previous section dealing with the structure of the ophiolitic rocks. Without the ophiolitic rocks, the presence of these major faults would probably not be suspected, and it seems reasonable to suppose that other similar faults without the characteristic ophiolitic slivers are present but have not been recognized.

Subunit 10b

Lithology and Distribution. Coarse conglomerate forms a minor component of the Cold Spring Pond Formation and has been recognized in two locations: 1) along the shore of Cold Spring Pond, and 2) in the bed of the West Salmon River southeast of Cold Spring Pond. These conglomerates are generally poorly sorted, vary from clast supported in the south to matrix supported in the north and are polymictic, containing principally felsic volcanic, greywacke and pelitic clasts and minor mafic volcanic fragments. The clasts are poorly sorted, and range in size from pebbles to boulders. The conglomerates are generally poorly bedded, but are

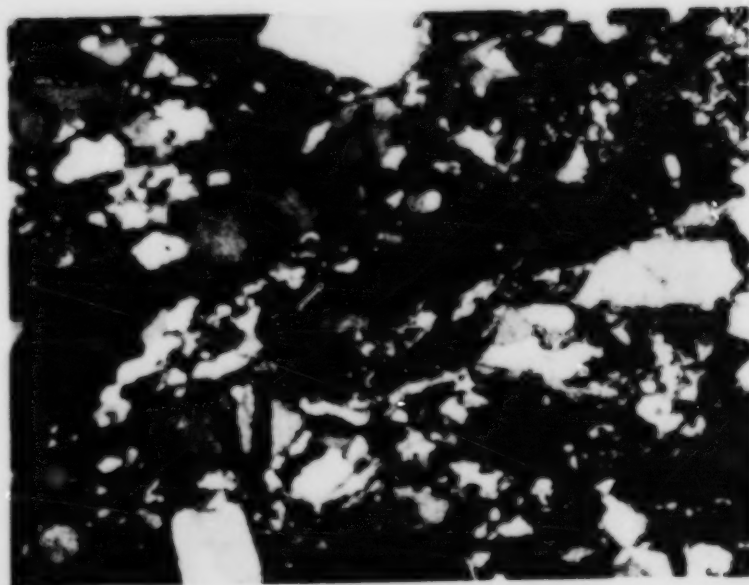


Plate 2-12: Photomicrograph of Cold Spring Pond Formation rhyolite illustrating severe cataclastic deformation involving comminution and recrystallization of matrix. Field of view is 6 mm.

interbedded locally with pebbly, graded greywacke and pelite, and are themselves slightly graded in some strata. These conglomerates appear to represent coarse grained lenses in the subunit 10a succession which, as previously noted, shows a tendency to coarsen to the south and west.

Structure and Contact Relations. Conglomerates of subunit 10b are stratigraphically conformable and probably gradational with clastic sedimentary rocks of subunit 10a which enclose them. These conglomerates contain the east-trending, principal fabric related to F_1 isoclinal folds in adjacent formations.

Subunit 10c

Lithology and Distribution. Subunit 10c consists of dark grey to black felsic volcanic rocks. These rocks make up approximately 25 percent of the Cold Spring Pond Formation. Their most widespread outcrop is a 1 to 1.5 km-wide linear belt, which can be traced from the North Salmon River to the area southeast of Great Burnt Lake, approximately 10 km to the southwest. Similar lithologies have been mapped by Colman-Sadd (1984) in the Cold Spring Pond (12A/1) map area to the southwest (see also Colman-Sadd and Swinden, 1983). These rocks also form lenses within the volcanoclastic succession of subunit 10a and, with conglomerates of subunit 10b, form the southern limit of the formation along the West Salmon River at the southern limit of mapping. Good exposures of typical lithologies which comprise subunit 10c can be seen in road cuts and quarries along the North Salmon Dam access road near the junction with the West Diversion access road and in the bed of the West Salmon River east of Cold Spring Pond.

The principal lithology of subunit 10c is highly siliceous 'cherty' crystal-lithic tuff. This rock is typically composed of an aphanitic groundmass, generally dark grey to black, but locally bluish or dark green on fresh surfaces, which weathers to a distinctive white clay. Variable amounts (from 2 to 25 percent) of crystals, crystal fragments and lapilli-sized lithic fragments are present in amounts ranging from 10 to 70 percent. Quartz is the most abundant phenocryst phase, occurring as somewhat rounded crystals and crystal fragments; it forms the largest (1 to 3 mm) and most abundant (60 to 80 percent) component of the lithic fragments. In most outcrops of subunit 10c, the resistant quartz grains stand out prominently on the weathered surface, giving the outcrops a distinctive knobby appearance. Feldspar crystals and crystal fragments were recognized in some but not all exposures; when present, they comprise from 20 to 45 percent of the clasts and average about half the size of the coexisting quartz. Lithic lapilli are not generally recognized in outcrop and seldom form more than 10 percent of the lithic assemblage. Where present, they are principally felsic volcanic rocks similar to the groundmass of the host. Rarely, flattened pumice is preserved.

Rocks of subunit 10c are locally hydrothermally altered with patchy development of chlorite and sericite and 1 to 5 percent disseminated pyrite.

In thin section, most specimens show evidence of severe cataclastic deformation which has destroyed virtually all original textures (Plate 2-12). All rocks examined consist of variable proportions of matrix, in which a likewise variably constituted, phenocryst-clast assemblage is set. The latter phases vary from less than 10 percent in some porphyritic

rhyolite tuffs to up to 70 percent in crystal-rich, crystal-lithic tuff. In the porphyritic rhyolitic tuffs, the principal phenocryst phase is quartz, ranging from 1 to 3 mm in diameter. It is typically well rounded, deeply embayed and isolated in a siliceous, aphanitic groundmass. A second clast type commonly observed consists of wispy, elongate lapilli from 1 to 3 mm across, consisting of biotite (60 percent), quartz and feldspar (35 percent) and minor sericite, chlorite and, locally, calcite. These are probably recrystallized lithic lapilli, either pelitic sedimentary rocks or perhaps cognate volcanic fragments that are somewhat more mafic than their rhyolite host. The groundmass typically comprises a granoblastic aggregate of quartz and feldspar with chlorite and sericite flakes ranging in abundance from 10 to 30 percent.

Crystal-lithic tuffs with up to 70 percent clasts are more commonly identified in thin section than in outcrop. These rocks consist of mixtures of quartz (40 to 60 percent), feldspar (40 to 60 percent) and lithic lapilli set in a sheared, recrystallized matrix of granoblastic quartz and feldspar with minor sericite and chlorite. In some specimens, micas comprise less than 5 percent of the groundmass. Quartz phenocrysts are commonly larger than feldspar, ranging in some specimens up to 6 mm. Where preserved in a recognizable form, they are generally embayed. Lithic lapilli consist of felsic rock fragments, now recrystallized to quartz, feldspar and minor sericite. Accessory phases in all rocks typically consist of apatite, zircon, sphene and opaque minerals.

The most prominent textures visible in most sections are those due to cataclastic deformation. Phenocrysts and lapilli are commonly shattered and extensively recrystallized both along grain boundaries and along the internal fracture surfaces. The groundmass is generally almost completely recrystallized; sericite and chlorite are strongly oriented in the plane of the principal foliation and wrap around the phenocrysts and lapilli to form augen. Coarse granoblastic quartz and sericite commonly form pressure shadows in the lee of the augen. Only minor feldspar alteration has taken place during this deformation. Most phenocrysts are only slightly saussuritized, although in some specimens almost complete alteration has occurred. Calcite is present locally as an alteration mineral after feldspar, along with chlorite-sericite-quartz, and is also found locally in calcite-chlorite-quartz-biotite veinlets, which generally parallel the main fabric.

The rocks that make up subunit 10c apparently comprise a felsic ash-flow tuff unit. No very coarse phases, indicative of proximity to an explosive vent, were found. In spite of relatively abundant lithic lapilli in some outcrops, no fragments of mafic volcanic rocks appear to be present, which is somewhat puzzling considering the spatial relationship between these lithologies throughout the central part of the Cold Spring Pond Formation.

Contact Relations. Contact relations between subunit 10c and adjacent lithologies appear to be relatively straightforward. The felsic volcanic rocks are interpreted to

be interbedded conformably with volcanoclastic sediments of subunit 10a and basalts of subunit 10d east of Great Burnt Lake and, as well, with conglomerates of subunit 10b near the West Salmon River. Intense folding precludes any determination of a uniform sense of facing in these rocks. A large lens of felsic volcanic rocks has been recognized northwest of Godaleich Pond. This may be a stratigraphic lens interbedded in the sequence; alternatively, it may be either a fault-bounded remnant in, or a window through, the volcanoclastic sequence exposed in a fold hinge.

At its southwestern end, the main belt of subunit 10c felsic volcanics is cut off by a series of faults which juxtapose it with ophiolitic rocks north of Cold Spring Pond.

Structure. The principal deformation visible in the rocks of subunit 10c is a moderate to strong penetrative cleavage, which generally strikes east-northeast to northeast with vertical to steep westward dips. This cleavage parallels that in adjacent units, which are related to F_1 isoclinal folding and, therefore, it is likewise attributed to a D_1 deformational episode. The distinctive feature of the deformation in subunit 10c is its intensely cataclastic nature, which is much more pronounced than in other lithologies in the Cold Spring Pond Formation. It is also more pronounced than in similar felsic volcanic rocks of the North Steady Pond Formation. The relative intensity of cataclasis in these rocks probably reflects at least two causes. First, the rocks of subunit 10c, especially those which contain relatively sparse amounts of chlorite, are more internally homogeneous than adjacent rocks and were therefore probably more brittle during deformation and more inclined to respond cataclastically to stress. Second, the felsic volcanic rocks of the North Steady Pond Formation (subunit 8d) are generally less deformed than those of the Cold Spring Pond Formation. Deformation was probably not sufficiently intense in the former to produce extensive cataclastic textures.

Subunit 10d—The North Salmon Dam Basalt

Definition and Distribution. The name North Salmon Dam Basalt is formally proposed for the continuous belt of pillow lava, massive basalt and minor hyaloclastic and pelitic volcanoclastic rocks that form the northeastern part of the Cold Spring Pond Formation. The basaltic rock outcrops in a 0.5 to 1 km-wide linear belt roughly centred on the North Salmon River and extending from this location approximately 7 km to the southwest and 10 km to the northeast. The proposed name is taken from the North Salmon Dam, a hydroelectric dam constructed across the North Salmon River. Good outcrops of the basalt are found in the area of this dam. The type section is designated as the dry bed of the North Salmon River east of the dam where the unit is exposed in an almost continuous section for approximately 600 m.

Lithology. Basaltic flow rocks of the North Salmon Dam Basalt comprise dominantly pillowed and massive flows, apparently in more or less equal proportions with lesser amounts of interbedded chloritic schist, probably derived from mafic tuffaceous rocks. Massive flows in outcrop are generally dark green, aphanitic and featureless except for local

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concentrations of minute feldspar phenocrysts and veinlets and patches of epidote and calcite. In some areas (e.g., in road cuts less than 1 km south of the North Salmon Dam) the massive flows are highly vesicular, containing up to 35 percent chlorite-calcite-epidote-filled amygdules. Autobrecciated massive flow, consisting of large (0.3 to 0.7 m) basalt blocks in a fine grained basaltic matrix is locally present. Pillows, generally 0.3 to 0.7 m across, slightly to moderately flattened, and locally variolitic were recognized in slightly less than half of the basalt outcrops. Like the massive flows, they commonly contain minute feldspar phenocrysts and chlorite-calcite-epidote-filled vesicles. Pillow rims are commonly epidotized and the pillows are cut by calcite- and epidote-filled veinlets. Pillow breccia was seen in one locality but does not appear to be widespread. Limestone or chert was observed interbedded with the volcanics or in the interstices between pillows, and the lavas are not variolitic.

Locally, the basalts exhibit intense chloritic alteration, accompanied by variable amounts of disseminated to massive sulphide mineralization. This feature will be further discussed in the chapter dealing with economic geology.

In thin section, the basalts are seen to consist mainly of a fine grained mosaic of feldspar, chlorite and amphibole (generally actinolite but with lesser amounts of hornblende). The groundmass is generally a fine grained to microcrystalline aggregate of feldspar (albite), amphibole and chlorite with subordinate amounts of sphene and minor epidote. Lath-shaped feldspar forms the principal phenocryst phase (5 to 35 percent of the rock), occurring as small, randomly oriented crystals from 0.05 to 2 mm long. Sparse, relict pyroxene phenocrysts, now mostly altered to amphibole-albite and chlorite are present. Actinolite is a common alteration mineral, both in the groundmass and as alteration of phenocrysts; it commonly forms pleochroic prismatic crystals, generally in association with calcite and epidote. Epidote is less common in thin section than is apparent in hand specimens. It typically forms isolated euhedral crystals in the groundmass and occurs with actinolite and calcite in veinlets and vesicle fillings. Chlorite is a common constituent of the microcrystalline groundmass but is less common in the coarser grained alteration assemblages in which amphibole predominates.

Vesicles are present in most basalts and they contain a wide variety of filling minerals. In some specimens, calcite is dominant, usually with only minor actinolite and epidote. Commonly, however, variable proportions of albite-chlorite-epidote are the dominant minerals with relatively minor (less than 25 percent) calcite.

Thin section examination of the chloritic schists that are commonly interbedded with the volcanic rocks reveals that they have a similar mineralogy to the more massive basalts. Albite-chlorite-amphibole are the principal phases present, and fragments of basalt are locally recognized. The chlorite and amphibole commonly show a preferred orientation that

is kinked by a later deformational event of sufficient intensity to locally develop a spaced cleavage. Sphene, epidote, apatite and opaques are accessory phases. The schists are compositionally very similar to the basalts but display no vestige of the igneous textures that characterize the latter. This, coupled with the presence of basaltic clasts and field observations of probable bedding features, indicates that they represent intervals of sedimentation between volcanic episodes, in which the sediments were derived from the local mafic volcanic pile.

Contact Relations. The main outcrop belt of the North Salmon Dam Basalt is interbedded to the east with volcanoclastic rocks and felsic volcanic rocks of the Cold Spring Pond Formation, except along a short section north of the North Salmon River. Here the eastern boundary fault of the formation cuts the eastern border of the basalt, juxtaposing it with arkose of the North Steady Pond Formation and gabbro of the Pipestone Pond Complex. Most of the western margin of this belt is cut by the western boundary fault of the Cold Spring Pond Formation, juxtaposing the basalts with metamorphic and igneous rocks of Units 11 and 13, respectively. This fault is marked by serpentinite slivers south of the North Salmon River and is exposed in road cuts on the hill immediately south of the North Salmon Dam.

Structure. The principal fabric in the North Salmon Dam Basalt is a northeast-to north-trending, moderate to strong, cleavage, which is steeply dipping and which is probably related to the F_1 isoclinal folding that characterizes the Cold Spring Pond Formation. North of the North Salmon River, this cleavage is refolded around moderately open, asymmetrical folds with steep eastern limbs, the axes of which are more or less parallel to those of the earlier isoclines. These F_1 folds are probably responsible for the kinking of oriented chlorite-amphibole bands and the development of spaced cleavage referred to in the previous section.

Metamorphism

Metamorphism in the Cold Spring Pond Formation is best displayed in the mafic volcanic rocks. In all areas, they contain a spilitic groundmass assemblage of chlorite-albite-actinolite-(calcite), indicating regional metamorphism in the lower to middle greenschist facies. Similar metamorphic assemblages are present in the coarser grained sedimentary rocks.

Correlation and Age

There is no direct evidence in the Great Burnt Lake area as to the age of the Cold Spring Pond Formation. It is unfossiliferous and is everywhere faulted against adjacent rock units. It differs from the volcanic-bearing formations of the Baie d'Espoir Group mainly in that it contains significant proportions of mafic volcanic rocks. The mafic volcanics are geochemically similar to rocks elsewhere in central Newfoundland, which range from Arenig to Caradocian.

UNIT 11

Definition and Distribution

Unit 11 includes metasedimentary and lesser amounts of metavolcanic rocks, which comprise the country rocks of the western igneous-metamorphic domain underlying the southwestern margin of the map area. These rocks occur in three separate parts of the present map area, and in each area, exhibit distinctive lithologies and structural style.

Subunit 11a

Lithology and Distribution. Subunit 11a underlies a linear, north- to south-trending belt approximately 5.5 km long and 200 to 400 m wide northeast of Great Burnt Lake. The rocks in this belt comprise medium-bedded, quartzite interbedded with lesser amounts of fine grained semipelitic material, which has been metamorphosed to garnet-mica-andalusite schist. Quartzite beds are commonly brown to dark grey and generally range from 5 to 20 cm thick, but locally attain thicknesses of up to 40 cm. The semipelitic intervals are commonly 5 to 10 cm thick, but in some exposures form the dominant rock type on outcrop scale. Garnet and staurolite porphyroblasts are locally present. The original lithologies exposed in this belt would probably have been very similar to those exposed in the Spruce Brook Formation (Unit 12) near Sitdown Pond.

Only one thin section of these rocks was examined. It was a sample of the quartzite, which consisted mainly of polygonized quartz grains (greater than 90 percent) and minor detrital plagioclase and biotite. Accessories are detrital zircon and opaque minerals (up to 2 percent). The quartzite is laminated on thin section scale. Fine laminae contain up to 50 percent intergrown biotite and chlorite; the chlorite locally forms radiating fibrous aggregates. Quartz and feldspar tend to be coarser grained than the platy minerals. Opaque minerals are abundant in the mafic-rich laminae.

Contact Relations. Subunit 11a maps out as a sinuous, linear raft within the deformed, porphyritic granite of Unit 13. The contacts were not seen in the field and it is not clear whether this belt forms a large xenolith or whether it is fault-bounded and structurally interleaved with the surrounding granite. Local development of migmatite near the contacts suggests the former interpretation. However, the belt appears to be devoid of any granitic dykes in its interior. Furthermore, the long, linear shape of this belt, which parallels the regional structural grain, as well as the prevalence of well-documented thrust faults, the traces of which also parallel this grain, argue for the latter possibility. In the absence of actual observations of the contact, the exact nature of the relationship between this unit and the surrounding rocks remains unresolved.

Structure and Metamorphism. Subunit 11a exhibits one strong penetrative foliation, defined by the alignment of biotite-chlorite-($\pm$ sericite), which strikes north to north-northwest, has steep to vertical dips and is axial planar to

tight, upright isoclinal folds. The first-generation isoclinal folds in subunit 11a are refolded by a second generation of gentle, open folds, the axes of which strike east-northeast to east and plunge gently to the east. Local development of a weak, east-trending, second cleavage in the sediments may be related to this episode of folding.

The presence of biotite, garnet and staurolite in pelitic rocks suggests that subunit 11a has been metamorphosed in the upper-greenschist- to lower-amphibolite facies.

Subunit 11b

Lithology and Distribution. A narrow band of metasedimentary rocks assigned to subunit 11b underlies an area approximately 2 km long and 100 m wide east of Great Burnt Lake, adjacent to the North Salmon Dam access road. These rocks are well exposed along the shoreline of the North Salmon River, immediately west of the North Salmon Dam, and in a road cut south of the North Salmon Dam. They comprise intensely sheared and deformed pelitic and semipelitic schists, some of which exhibit incipient gneissic banding. Along the North Salmon River, some exposures contain pink feldspar porphyroblasts from 0.5 to 2 cm across, which overgrow the foliation; the host rocks are dense black amphibolites which may represent either sedimentary or mafic volcanic protoliths. Garnet, staurolite and andalusite(?) are locally developed in the pelitic rocks.

In thin section, the pelitic and semipelitic rocks consist principally of polygonized quartz and lesser amounts of detrital feldspar grains. Green biotite is the most important mafic mineral, forming up to 15 percent of the rock. Detrital zircon and granular aggregates of epidote are relatively abundant (2 to 4 percent) and garnet is also present. Hornblende and staurolite locally overgrow the earlier formed minerals.

Subunit 11b is generally poorly exposed, so that its full extent is uncertain. It is presently thought to underlie a relatively small area as similar lithologies do not outcrop along strike southeast of Great Burnt Lake. However, outcrop is generally scarce in this area and it is possible that these rocks may underlie a continuous belt adjacent to the western boundary fault of the Baie d'Espoir Group, connecting to the south with Unit 11c southeast of Great Burnt Lake.

Contact Relations. To the east, subunit 11b is juxtaposed with, greenschist-facies pillow basalts of the Cold Spring Pond Formation along a steeply west-dipping thrust fault. This fault is exposed in a road cut immediately south of the North Salmon Dam where a small body of serpentized peridotite has been emplaced along it. The western contact with foliated granite was not seen in the field, but is assumed to be intrusive as flattened xenoliths lithologically similar to subunit 11b lithologies occur within the granite elsewhere in the area.

Structure and Metamorphism. Pelitic rocks of subunit 11b carry a single, intense, penetrative schistosity which trends

PIPESTONE POND AREA

slightly east of north and dips subvertically. The schistosity is defined by the alignment of biotite and chlorite. Garnet and staurolite appear in hand specimen to overgrow this foliation and therefore probably postdate the peak of principal deformation.

Subunit 11c

Lithology and Distribution. Subunit 11c underlies the southwestern part of the map area between Great Burnt Lake and Cold Spring Pond. Here metasedimentary and metavolcanic rocks form the dominant lithologies north and west of the Cold Spring Pond Formation and extend westward to the limits of the present mapping. The principal rock types are micaceous, garnetiferous psammitic rocks intercalated with lesser amounts of semipelitic rocks. All have been intensely deformed (Plate 2-13); in some areas, original sedimentary features can still be recognized, but not uncommonly, the deformation is accompanied by migmatization. Recognizable volcanic rocks are rare, but northwest of Cold Spring Pond, sericitic schists with quartz phenocrysts and massive, white to buff, siliceous rocks that may represent a felsic volcanic protolith outcrop in a series of low hillocks.



Plate 2-13: *Highly deformed and migmatized metasedimentary rocks of subunit 11c northwest of Cold Spring Pond.*

Contact Relations. In the area north and northeast of Cold Spring Pond, subunit 11c is interpreted to be in fault contact with both the Cold Spring Pond Formation and with a large ophiolitic fragment, although these faults are not exposed.

Structure and Metamorphism. Tight polyphase folding is evident in many outcrops (Plate 2-13), and the development of migmatite is relatively widespread. In recognizable metasedimentary rocks, there is commonly a steeply dipping principal penetrative foliation, which strikes roughly east to west north of Cold Spring Pond and northeasterly in the area

southeast of Great Burnt Lake. This fabric is axial planar to isoclinal folds and is probably equivalent to the main fabric in subunits 11a and 11b.

Development of the principal foliation was accompanied by growth of biotite-garnet-staurolite and possibly hornblende. This suggests a regional metamorphism in the lower-amphibolite facies, although local production of migmatite suggests that metamorphic conditions locally approached that necessary for incipient partial melting.

Correlation and Age

The age of the rocks comprising Unit 11 is not known with any degree of certainty. The rocks constitute the eastern extremity of an extensive metasedimentary, metavolcanic and migmatitic terrane, which outcrops throughout south-central Newfoundland (e.g., Units 4a,b of Williams, 1970; Unit 3b of Kean and Jayasinghe, 1980; Unit 8 of Kean, 1983; Units 1-3 of Dickson, 1982; Unit Omc of Hibbard, 1983; Units 16-18 of Colman-Sadd and Swinden, 1984b). The rocks of subunit 11a are lithologically similar to, and may be correlative with, the Spruce Brook Formation. Similar rocks outcrop west of the present map area and their similarity to rocks of the Spruce Brook Formation has been noted by Colman-Sadd (1985b).

Rocks of subunit 11b are too poorly exposed to allow a definitive correlation with other areas. They are, however, lithologically similar to rocks to the west of the present area which have been assigned an Early to Middle Ordovician age by various workers (see citations in previous paragraph).

Subunit 11c displays many lithological similarities with the Baie d'Espoir Group. For example, in the area of the west diversion channel, it includes dense, purplish sandstones similar to Salmon River Dam Formation lithologies and north of Cold Spring Pond, felsic volcanics similar to lithologies in the Isle Galet and North Steady Pond formations are present. Accordingly, it is tentatively correlated with the Baie d'Espoir Group and assigned an Early to Middle Ordovician age.

SPRUCE BROOK FORMATION (UNIT 12)

Definition and Distribution

The Spruce Brook Formation was defined in the Burnt Hill (2D/5) map area by Colman-Sadd (1985a) to include a sequence of turbiditic quartzose sandstone and pelitic material occurring west of the western boundary faults of the Coy Pond-Great Bend Ophiolite complexes. The full extent of the Spruce Brook Formation as presently defined was shown by Colman-Sadd and Swinden (1982, their Figure 3), who subdivided the sequence into two informally named members, the Spruce Brook and Miguels Brook units. The Spruce Brook Formation underlies parts of the Burnt Hill (2D/5), Miguels Lake (2D/12) and Great Burnt Lake (12A/8) map areas.

In the Great Burnt Lake map area, the Spruce Brook Formation underlies a semicircular area of approximately 80 km extending from the eastern boundary fault of the Pipestone Pond Complex eastward to the map boundary.

Lithology

The Spruce Brook Formation in the Pipestone Pond area consists mainly of mature, quartzose sandstone and subordinate amounts of intercalated pelitic rocks. The sandstones, where relatively unmetamorphosed, are white to grey weathering, massive and thick bedded (up to 1 m). Finer grained pelitic layers, up to 4 cm thick, are locally intercalated with the sandstones. The pelitic beds commonly comprise less than 20 percent of any outcrop. Outcrops tend to be small and lichen covered, and do not commonly exhibit sedimentary structures, although local examples of graded bedding, scour and fill, and possible load structures are present in exposures along the south flowing brook that drains Sitdown Pond.

The protolith for all of the Spruce Brook Formation in the present area is represented by lower-greenschist-facies rocks in the area south and southwest of Sitdown Pond. In this area, well-washed quartz sandstones consisting of 85 to 95 percent rounded, detrital quartz grains up to 1 mm in diameter contain minor small laths and flakes of sericite and chlorite. The platy minerals are commonly oriented in the plane of the principal foliation. Interbedded pelitic beds contain proportionally more chlorite-sericite and less quartz, but otherwise seem to reflect the same provenance. Spinel, zircon and opaque minerals (mainly magnetite) are ubiquitous accessories. The metamorphism of this unit has produced mineralogical changes across the area, which are more fully documented below.

Contact Relations

Along most of its western boundary, the Spruce Brook Formation is in fault contact with ultramafic rocks of the Pipestone Pond Complex. This fault is presently interpreted to be a thrust fault along which the ophiolitic sequence has overridden the structurally underlying sediments (Colman-Sadd and Swinden, 1984a). The ultramafic rocks are pinched out along this fault system near the eastern edge of the map area, and in this region sedimentary and felsic volcanic rocks of the North Steady Pond Formation are juxtaposed with the Spruce Brook Formation.

Structure

The Spruce Brook Formation normally contains a strong penetrative foliation, although this may not be obvious in outcrops consisting only of quartzose sandstone. In relatively pelitic lithologies, the foliation is well developed; in more psammitic lithologies, it is commonly evidenced in thin section by alignment of sericite, chlorite and biotite. Folds related to this S_1 fabric were not seen in the present area, although Colman-Sadd (1985b) has documented tight, steeply

southwest-plunging, folds in the Burnt Hill area, which appear to be related to S_1 . East of Sitdown Pond, the S_1 foliation is folded around tight, north to south-trending folds, which plunge steeply to the south and have an associated, weak cleavage striking $340^\circ/010^\circ$ and dipping moderately to steeply westward. This cleavage is represented in thin section by kinking of previously aligned biotite and sericite grains.

Metamorphism

The Spruce Brook Formation exhibits rapid changes in metamorphic grade, particularly in the adjacent Burnt Hill area to the east where a complete progression from greenschist facies to migmatitized upper-amphibolite facies has been documented by Colman-Sadd (1985a) and Colman-Sadd and Swinden (1984a). A similar, although less complete, prograde succession is present in the Pipestone Pond area ranging from greenschist (chlorite + sericite) to lower-amphibolite (andalusite + garnet $\pm$ staurolite) facies.

As previously noted, the lowest grade rocks in the Spruce Brook Formation occur in the southeast near the North Salmon River. In this area, the metamorphic assemblage is lower greenschist, characterized by quartz-sericite-chlorite. Increasing eastward metamorphism is first evidenced by the appearance of prograde biotite defining the biotite isograd shown in Figure 2-1. The biotite is locally accompanied by staurolite. Further increase in the metamorphic grade results in the appearance of large (4 to 5 mm) andalusite crystals, generally accompanied by biotite, muscovite, staurolite and locally garnet in rocks of appropriate composition (Plate 2-14A,B). The andalusite isograd shown in Figure 2-1 is defined by the first appearance of andalusite; rocks of this metamorphic grade (probably lower-amphibolite facies) comprise more than half of the Spruce Brook Formation exposure in the Pipestone Pond area.

Growth of andalusite, staurolite and garnet both preceded and accompanied the D_1 deformational event as these minerals both overgrew the S_1 fabric defined by sericite-biotite alignment and locally form augen within it.

Static biotite and sericite growth continued past the D_1 deformational event in all areas and both occur locally as randomly oriented laths and flakes cutting the S_1 fabric. Locally, they are crenulated by a later event, probably the D_2 deformation previously noted.

Minor retrograde chlorite is present in some specimens replacing staurolite and biotite.

Correlation and Age

There is no direct evidence in the Pipestone Pond area as to the age of the Spruce Brook Formation. However, brachiopods have been recovered from two localities in this formation in the Miguels Lake (2D/12) map area (Colman-Sadd and Swinden, 1982, 1984a). The host lithology is a grey,

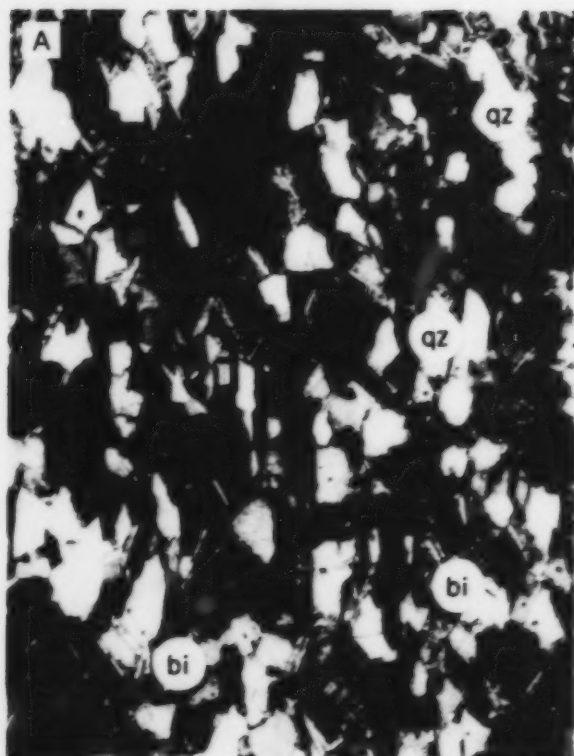


Plate 2-14: Metasedimentary rocks of the Spruce Brook Formation. In A, from the biotite zone, quartz (qz) and minor mafic minerals are overgrown by biotite (bi). In B, retrogressed staurolite (st) and garnet (g) coexist with biotite in semipelitic rocks. Field of view is 6 mm.

calcareous sedimentary breccia and the contained fauna indicates a Late Arenig-Caradoc (probably Llanvirn) age (Colman-Sadd and Swinden, 1984a).

Because it is fault bounded on all sides, the Spruce Brook Formation cannot be directly correlated with adjacent units in central and eastern Newfoundland. Based on the faunal evidence, it would appear to be the same age as the volcanic-sedimentary sequences of the Baie d'Espoir Group and Cold Spring Pond Formation, but it contains considerably more mature sandstones than these formations and lacks their ubiquitous volcanic detritus (a more detailed comparison of the two rock types with pointcounts of representative samples can be found in Colman-Sadd and Swinden, 1984a). It is lithologically very similar to the Gander Group to the east (Blackwood, 1982), but the two are 25 km apart in a direction perpendicular to strike and no direct stratigraphic correlation is possible.

Similar lithologies west of the present map area have recently been mapped by Colman-Sadd (1985b) who

suggested that they are probably related to the Spruce Brook Formation.

UNIT 13

Definition and Distribution

Unit 13 consists of variably deformed, locally gneissic, granitoid rocks that outcrop in the western part of the map area. The eastern boundary of Unit 13 marks the western limit of geological interest with respect to the objectives of the present program, and this unit was therefore not investigated in detail. The deformed granitoid rocks can be traced along a north to north-northeast trending belt up to 1.5 km wide, and extending from the area west of Gulp Pond southward to the area southeast of Great Burnt Lake (Williams, 1970; Colman-Sadd and Swinden, 1982, 1984b).

Lithology

Unit 13 consists mainly of variably foliated porphyritic (megacrystic) granite. The dominant characteristic in outcrop is the abundant pink to white feldspar phenocrysts, comprising from 15 to 60 percent of the rock and ranging in size from less than 0.5 cm to greater than 4 cm. Smaller quartz eyes (generally less than 3 mm) are commonly present and locally exhibit a bluish tinge. Biotite and chlorite, ranging in content from 5 to 20 percent, make up the mafic component in most outcrops. In the more intensely deformed localities, mafic minerals are coarsely crystalline and segregated into lenticular masses, 0.5 to 3 cm long, aligned in the plane of the foliation.

Amphibolitic and metasedimentary inclusions are locally present, and are commonly intensely flattened in the plane of the foliation (Plate 2-15). The inclusions are commonly recrystallized to the extent that the protolith can no longer be recognized. Outcrops of migmatite are sporadically present throughout Unit 13, but do not appear to form a coherent map unit within it. Fine grained pegmatites are generally associated with these migmatites, and are also locally present in the granite as injections along the foliation.



Plate 2-15: Deformed porphyritic granite of Unit 13 containing intensely flattened semipelitic inclusions.

Granitoid rocks of Unit 13 are seen in thin section to have been affected by intense cataclastic deformation and accompanying recrystallization. The larger feldspar phenocrysts are cataclastically deformed and broken microcline, which commonly contains exsolved albite forming perthitic intergrowths. The groundmass contains up to 40 percent plagioclase (An_{60}), which is variably (locally completely) saussuritized and locally displays myrmekitic intergrowths. Quartz grains are generally granoblastic and range from 0.5 to 1.5 mm, although larger phenocrysts are locally present.

Biotite is the principal mafic mineral. It occurs as dark green to brown flakes which are commonly segregated into foliae in the plane of the main deformation. The biotite is locally chloritized and commonly accompanied by syn- and post-kinematic muscovite, epidote and, rarely, garnet. Early-formed microcline phenocrysts form augen within biotite lamellae. Accessories include opaque minerals, zircon and sphene.

Contact Relations

The eastern contact of Unit 13, where it is in contact with the Cold Spring Pond Formation and the Pipestone Pond Complex, is commonly a high-angle reverse fault marked by a zone of intense shearing and the local presence of serpentinite slivers (e.g., immediately south of the North Salmon Dam). The presence of other fault-bounded ultramafic fragments of probable ophiolitic derivation west of the boundary fault northeast of Great Burnt Lake, suggests that similar faults may be present within the body of Unit 13, but due to poor exposure, the faults can only be mapped in detail where they bring up underlying exotic lithologies. The western contact of Unit 13 has not been observed in the field. Recent mapping by Colman-Sadd (1985b) indicates that rocks to the west comprise mainly equigranular, garnetiferous leucogranites although the contact relations have not been observed. The contact between Unit 13 and the metasedimentary rocks of Unit 11 is not exposed. However, inclusions in the granitoid rocks are lithologically similar to the metasediments, therefore it seems likely that the contact is intrusive.

Structure

Unit 13 displays a moderate to intense, north-northwest to north-northeast-trending, steeply dipping, foliation. The foliation is defined by stretching of quartz and feldspar, by parallel growth of biotite-chlorite in discrete lamellae in the plane of the foliation and by cataclastic-crush zones. The deformation is inhomogeneously developed, and the intensity of deformation bears no obvious specific relationship to mappable faults or other geological features, although it is generally parallel to the western boundary fault of the Cold Spring Pond Formation.

PIPESTONE POND AREA

Correlation and Age

There is no direct evidence in the Great Burnt Lake area

for the age of Unit 13. It contains inclusions of Unit 11 and therefore is probably not older than Early to Middle Ordovician.

CHAPTER 3

GEOCHEMISTRY

GENERAL STATEMENT

Representative suites of all igneous rocks in the Pipestone Pond Complex, the Cold Spring Pond and North Steady Pond formations have been analyzed for major- and trace-element concentrations. This chapter summarizes the results of these analyses.

In the first section, the various units of the Pipestone Pond Complex are considered. Following this, the chemistry of mafic extrusive rocks of the Cold Spring Pond Formation will be described and comparisons made with those of the Pipestone Pond Complex. Finally, there is a brief description of felsic volcanic rocks of the Cold Spring Pond and North Steady Pond formations.

Only summaries and averages of geochemical data, with major elements recalculated on an anhydrous basis, are presented in this chapter. Major element data on all diagrams have been recalculated on an anhydrous basis. Summaries of analytical methods, precision and accuracy are given in Appendix 1. Complete raw data sets are presented in Appendix 2. Throughout this chapter, the notation FeO* refers to total iron as FeO.

PIPESTONE POND COMPLEX

Ultramafic Rocks

Harzburgite

Results of geochemical analyses of three harzburgite samples are presented in Table 3-1. The analyzed samples were selected from hand specimens that showed the least evidence of serpentinization.

Typical ophiolitic harzburgites commonly exhibit a relatively restricted range of compositions (Coleman, 1977). On the AFM triangular diagram (Figure 3-1), samples from the Pipestone Pond Complex plot within compositional fields of typical harzburgite, and on the $\text{SiO}_2\text{--FeO}^*/\text{FeO}^*+\text{MgO}$ binary plot (Figure 3-2) they also plot close to the average metamorphic harzburgite composition.

Pyroxenite

In the Pipestone Pond Complex, cumulate pyroxenite forms the stratigraphically lowest unit of the cumulate complex. Ultramafic cumulates in other ophiolites show a wide range in composition, reflecting varying proportions of cumulate ortho- and clinopyroxene, olivine and plagioclase. Rock types may range from cumulate peridotite to olivine gabbro (Coleman, 1977). Results of four analyses of cumulate pyroxenite from the Pipestone Pond Complex are summarized

TABLE 3-1. Summary of geochemical analyses of harzburgites from the Pipestone Pond Complex.

	Pipestone Pond Complex (n=3)			
	Mean	St. Dev.	Min.	Max.
SiO_2	42.1	1.7	40.2	43.1
Al_2O_3	0.39	0.21	0.71	0.59
Fe_2O_3	3.87	0.49	3.36	4.33
FeO	4.20	0.68	3.43	4.65
MgO	48.53	2.58	46.78	51.49
CaO	0.59	0.41	0.16	0.98
Na_2O	0.06	0.01	0.06	0.07
K_2O	0.01	0.01	0.01	0.02
TiO_2	0.08	0.05	0.03	0.13
MnO	0.13	0.02	0.11	0.15
P_2O_5	0.01	0.00	0.01	0.01
Cu	11	5	5	14
Zn	32	2	30	33
Ni	2217	110	2130	2340
Cr	538	207	378	772

Notes: Data have been recalculated on an anhydrous basis. Values for oxides in weight percent; values for the trace elements, Cu, Zn, Ni and Cr in parts per million (ppm). Number of samples given by n; St. Dev. = standard deviation; Min. = minimum value; Max. = maximum value.

in Table 3-2. Compared to typical ultramafic cumulates in other ophiolites, those in the Pipestone Pond Complex are strongly enriched in SiO_2 and CaO and depleted in MgO. These anomalous compositions reflect the predominance of cumulate clinopyroxene in the rocks. Orthopyroxene and olivine occur only as rare cumulate crystals and interstitial material and the bulk rock compositions summarized in Table 3-2 approach the composition of augite.

The Pipestone Pond Complex pyroxenites plot mainly near the ultramafic end of the cumulate field on the AFM diagram (Figure 3-1). On the $\text{SiO}_2\text{--FeO}^*/\text{FeO}^*+\text{MgO}$ diagram (Figure 3-2), their relatively SiO_2 -rich compositions cause them to plot outside the field of typical ultramafic and mafic cumulates. These departures reflect the predominance of clinopyroxene and the relative scarcity of olivine, orthopyroxene and feldspar in the rocks.

Mafic to Felsic Intrusive and Extrusive Rocks

Four intrusive lithologies associated with the ophiolitic succession were recognized in the field: 1) gabbro; 2) diabase dykes and fine grained, mafic intrusive lenses; 3) plagiogranite dykes and pods and 4) very coarse grained,

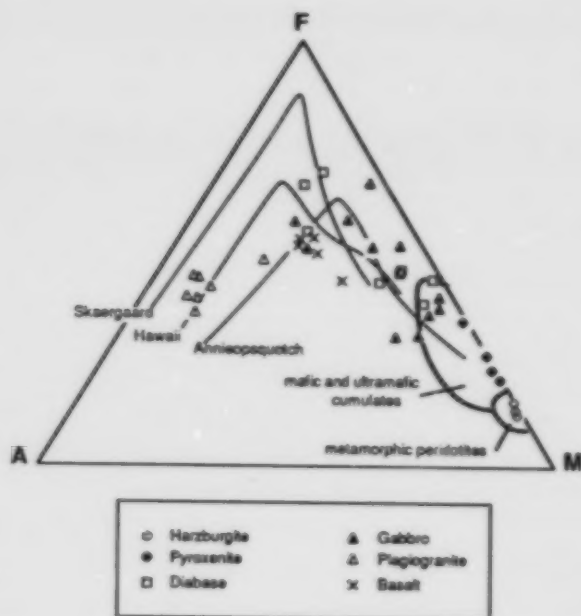


Figure 3-1: AFM diagram for igneous rocks of the Pipestone Pond Complex. Fields for ultramafic and mafic cumulate rocks and the Skaergaard trend from Coleman (1977) and Annieopsquitch trend from Dunning (1984).

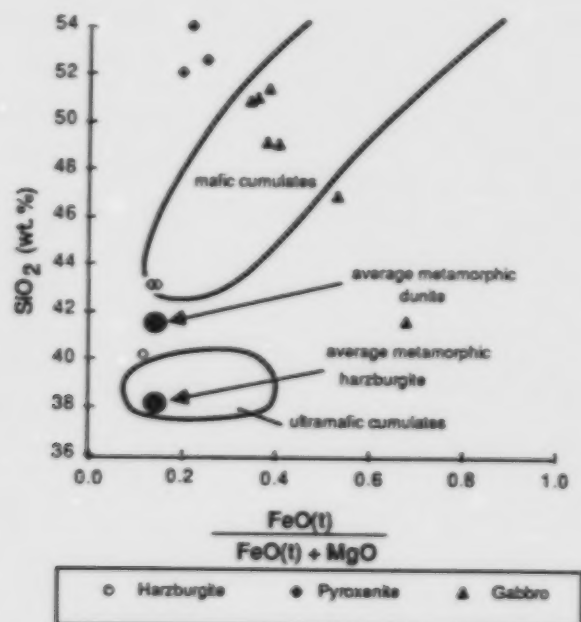


Figure 3-2: SiO_2 - $\text{FeO}^*/\text{FeO}^*+\text{MgO}$ plot for mafic and ultramafic plutonic rocks of the Pipestone Pond Complex. Fields from Coleman (1977).

TABLE 3-2. Summary of geochemical analyses of pyroxenite from the Pipestone Pond Complex

Pipestone Pond Complex (n=4)				
	Mean	St.Dev.	Min.	Max.
SiO_2	53.7	2.0	52.0	56.4
Al_2O_3	2.17	1.00	0.88	3.30
Fe_2O_3	1.03	0.49	0.39	1.55
FeO	6.20	2.38	4.25	9.65
MgO	20.21	1.51	18.86	21.99
CaO	16.31	3.09	11.99	18.85
Na_2O	0.09	0.02	0.07	0.11
K_2O	0.02	0.01	0.01	0.03
TiO_2	0.09	0.03	0.06	0.12
MnO	0.16	0.04	0.11	0.19
P_2O_5	0.01	0.00	0.01	0.01
Cu	181	128	11	281
Zn	32	10	24	46
Ni	220	47	167	264
Cr	1594	725	726	2498

Notes: Data have been recalculated on an anhydrous basis. Values for oxides in weight percent; values for the trace elements, Cu, Zn, Ni and Cr in ppm. Number of samples given by n; St. Dev. = standard deviation; Min. = minimum value; Max. = maximum value.

pegmatitic gabbro dykes. Geochemical data are reported for representative samples of the first three types; the pegmatitic gabbro is too coarse grained to allow a meaningful chemical analysis of normal hand-specimen-sized samples. Pillow basalts and fine grained basaltic flows form the upper stratigraphic levels of the ophiolite south of Pipestone Pond.

Gabbro

Results of analyses of eleven coarse grained, equigranular to porphyritic gabbro samples are presented in Table 3-3. The rocks range from slightly quartz- to slightly olivine-normative.

Five samples from the Pipestone Pond Complex plot in and near the field of mafic cumulates on the AFM diagram (Figure 3-1) and the very high $\text{Mg}\#$ in these samples (>0.70 , Table A9) further indicates the accumulation of pyroxene and/or olivine. However, six samples with lower $\text{Mg}\#$ (<0.63 , Table A9) plot outside this field, towards more iron-rich compositions. It is possible that these samples are part of an iron-enrichment trend and more closely approximate liquid compositions than their less iron-rich counterparts. All samples plot in or near the field of mafic cumulates on the SiO_2 - $\text{FeO}^*/\text{FeO}^*+\text{MgO}$ diagram (Figure 3-2); the single exception (Sample 2140332) is particularly SiO_2 poor and iron rich due to abundant magnetite and, therefore, plots outside the normal igneous rock fields.

TABLE 3-3. Summary of geochemical analyses and CIPW normative minerals of gabbro from the Pipestone Pond Complex.

Pipestone Pond Complex (n=11)				
	Mean	St.Dev.	Min.	Max.
SiO ₂	51.8	5.6	41.6	62.3
Al ₂ O ₃	16.01	2.40	10.26	18.98
Fe ₂ O ₃	2.28	1.92	0.83	7.77
FeO	7.74	2.21	4.90	11.16
MgO	8.96	3.26	4.11	14.83
CaO	10.71	3.55	3.50	14.87
Na ₂ O	1.92	1.30	0.45	4.69
K ₂ O	0.13	0.11	0.02	0.31
TiO ₂	0.28	0.20	0.04	0.70
MnO	0.20	0.04	0.14	0.26
P ₂ O ₅	0.01	0.00	0.01	0.01
Cu	78	87	19	315
Zn	58	22	8	90
Ni	36	24	5	80
Rb	6	2	3	9
Sr	81	39	21	161
Y	4	4	1	12
Zr	5	3	1	10
Nb	1	1	1	4
Ga	8	6	1	26
Ba	51	30	10	108
V	335	356	107	1382
Cr	139	120	5	380
Normative Minerals				
Quartz	4.01	6.27	0.00	19.46
Orthoclase	0.75	0.67	0.12	1.85
Albite	16.26	11.03	3.84	39.71
Anorthite	34.50	9.51	17.34	45.87
Diopside	15.04	9.33	0.00	30.73
Hypersthene	20.95	7.81	7.33	30.83
Olivine	4.55	4.96	0.00	13.46
Magnetite	3.30	2.78	1.21	11.27
Corundum	0.03	0.03	0.00	0.09
Ilmenite	0.74	0.71	0.08	2.55
Apatite	0.02	0.00	0.02	0.03

Notes: Data recalculated on an anhydrous basis. Values for oxides in weight percent; values for the trace elements, Cu to Cr in ppm. Number of samples given by n; St. Dev. = standard deviation; Min. = minimum value; Max. = maximum value.

Diabase

Eight samples of fine grained intrusive rocks were analyzed (Table 3-4), including samples from dykes cutting the gabbro in the northern part of the complex and from the diabase lenses which form the top of the intrusive section in the Jamieson Hills.

TABLE 3-4: Summary of geochemical analyses of diabase from the Pipestone Pond Complex.

Pipestone Pond Complex (n=8)				
	Mean	St.Dev.	Min.	Max.
SiO ₂	53.7	4.4	49.4	63.4
Al ₂ O ₃	14.38	1.65	11.88	17.28
Fe ₂ O ₃	2.34	0.83	1.42	4.01
FeO	8.56	2.62	3.63	11.72
MgO	8.67	4.06	1.99	13.16
CaO	9.43	2.74	5.42	14.10
Na ₂ O	2.01	1.29	0.44	4.79
K ₂ O	0.13	0.17	0.03	0.55
TiO ₂	0.49	0.29	0.15	1.03
MnO	0.22	0.06	0.12	0.30
P ₂ O ₅	0.04	0.03	0.01	0.08
Cu	63	50	17	173
Zn	77	49	33	182
Ni	75	69	1	186
Rb	4	2	2	8
Sr	94	42	49	157
Y	9	7	2	22
Zr	11	7	3	27
Nb	2	1	1	4
Ga	10	11	1	35
Ba	182	278	39	832
V	318	109	147	500
Cr	353	348	5	882
La	0.1	0.1	0.03	0.3
Ce	3.5	3.9	0.6	7.9
Nd	0.3	0.3	0.06	0.7
Sm	0.1	0.1	0.01	0.2
Eu	0.5	0.3	0.3	0.8
Gd	0.3	0.2	0.1	0.4
Tb	0.2	0.2	0.1	0.5
Er	0.2	0.1	0.03	0.3
Yb	0.1	0.1	0.01	0.2

Notes: Data recalculated on an anhydrous basis. Values for oxides in weight percent; values for the trace elements, Cu to Cr in ppm; values for the rare-earth elements La to Yb in ppm. Number of samples given by n; St. Dev. = standard deviation; Min. = minimum value; Max. = maximum value.

With one exception, the diabase samples have Mg# < 0.68, compatible with petrographic evidence that these rocks are generally not cumulates and represent a range of liquid compositions. On the AFM diagram (Figure 3-1), most samples are only slightly more iron rich than the cumulate gabbros. Three samples plot well above the cumulate field and may be part of the iron-enrichment trend, which includes the most fractionated gabbro samples. If this trend results from crystal fractionation, the latter samples may represent more evolved liquid compositions.

PIPESTONE POND AREA

Plagiogranite

Seven analyses of plagiogranite, including samples from dykes and from lenses in the upper part of the cumulate complex, are summarized in Table 3-5. Normative feldspar compositions of four samples are characteristic of trondhjemite and three are characteristic of tonalite (Figure 3.3A).

TABLE 3-5. Summary of geochemical analyses and CIPW normative minerals of plagiogranite from the Pipestone Pond Complex.

Pipestone Pond Complex (n=7)				
	Mean	St.Dev.	Min.	Max.
SiO ₂	73.1	2.6	67.3	75.0
Al ₂ O ₃	13.97	2.00	12.87	18.44
Fe ₂ O ₃	0.65	0.32	0.27	1.20
FeO	3.81	1.06	2.54	5.19
MgO	1.28	0.53	0.75	2.10
CaO	2.27	1.03	1.14	3.75
Na ₂ O	4.58	1.38	2.93	7.18
K ₂ O	0.35	0.21	0.14	0.65
TiO ₂	0.34	0.15	0.19	0.67
MnO	0.09	0.03	0.05	0.13
P ₂ O ₅	0.06	0.05	0.02	0.17
Cu	16	9	6	30
Zn	44	20	24	75
Ni	1	0	1	1
Rb	5	5	2	15
Sr	103	51	65	213
Y	32	18	15	69
Zr	45	57	12	172
Nb	1	1	1	4
Ga	10	5	5	19
Ba	115	86	46	284
V	34	17	14	57
Cr	5	0	5	5

Normative Minerals

Quartz	35.66	10.43	13.44	45.56
Orthoclase	2.03	1.21	0.85	3.83
Albite	38.47	11.08	24.82	58.82
Anorthite	10.87	5.28	5.26	18.39
Diopside	0.00	0.00	0.00	0.00
Hypersthene	9.18	3.20	5.63	13.63
Olivine	0.00	0.00	0.00	0.00
Magnetite	0.95	0.47	0.38	1.75
Corundum	0.00	0.00	0.00	0.00
Ilmenite	0.64	0.28	0.37	1.23
Apatite	0.14	0.12	0.05	0.39

Notes: Data recalculated on an anhydrous basis. Values for oxides in weight percent; values for the trace elements, Cu to Cr in ppm. Number of samples given by n; St. Dev. = standard deviation; Min. = minimum value. Max. = maximum value.

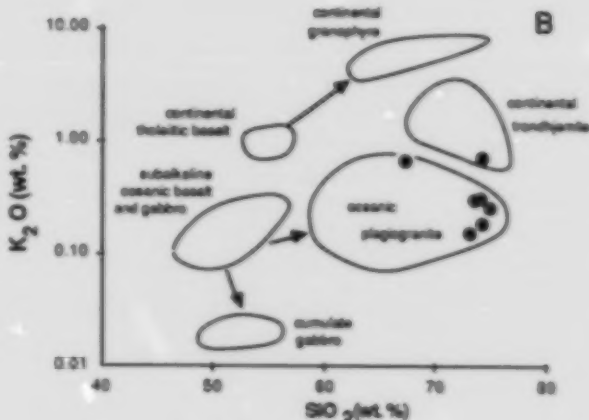
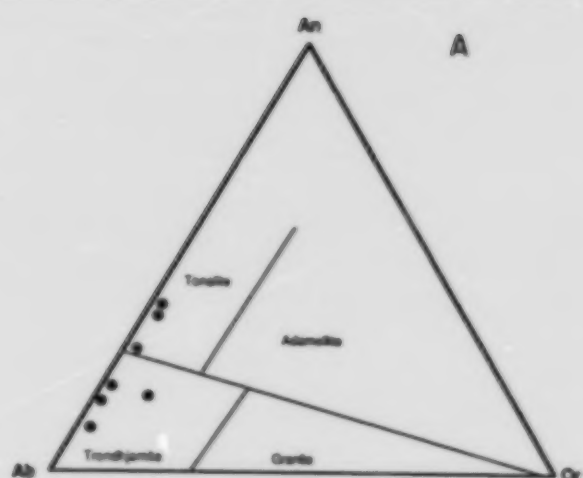


Figure 3-3: Classification of plagiogranites. A: triangular diagram showing classification of plagiogranites according to their normative feldspar compositions. Classification after O'Connor (1965). Solid circles are plagiogranites from the Pipestone Pond Complex. B: SiO₂-K₂O diagram illustrating the distinction between oceanic and continental plagiogranites. Fields from Coleman (1977). Pipestone Pond Complex plagiogranites plot mainly in the field of oceanic plagiogranites.

On the AFM diagram (Figure 3-1), the plagiogranites plot along an alkali enrichment trend similar to the most evolved rocks of the Hawaiian and Skaergaard trends.

Coleman (1977) noted that low K₂O contents are characteristic of oceanic plagiogranites and serve to distinguish them from their continental counterparts. This is illustrated by the K₂O-SiO₂ diagram (Figure 3-3B), in which most samples from the Pipestone Pond Complex plot within or near the field of oceanic plagiogranites.

Basalt

Five samples of basaltic rocks were analyzed during the present study the results of which are summarized in Table 3-6. In all cases, samples were taken from the interiors of pillows, and an attempt was made to analyze samples which were vesicle- and vein-poor and which showed a minimum of alteration.

TABLE 3-6. Summary of geochemical analyses of basalt from the Pipestone Pond Complex.

Pipestone Pond Complex (n=5)				
	Mean	St.Dev.	Min.	Max.
SiO ₂	50.9	1.10	49.4	52.3
Al ₂ O ₃	14.85	0.83	13.75	15.91
Fe ₂ O ₃	3.60	2.59	0.61	7.95
FeO	9.05	1.43	7.03	10.68
MgO	6.63	0.67	5.89	7.64
CaO	7.94	1.84	5.67	10.50
Na ₂ O	4.15	0.21	3.87	4.37
K ₂ O	1.01	0.69	0.12	1.66
TiO ₂	1.49	0.19	1.19	1.70
MnO	0.22	0.03	0.19	0.26
P ₂ O ₅	0.13	0.02	0.10	0.15
Cu	48	14	31	65
Zn	88	18	61	106
Ni	56	17	42	86
Rb	20	12	7	34
Sr	91	30	48	127
Y	40	6	31	46
Zr	78	12	59	88
Nb	4	2	2	7
Ga	10	2	8	13
Ba	189	92	88	337
V	395	69	309	477
Cr	143	85	49	276
La	2.2	1.3	0.4	3.2
Ce	9.25	1.85	6.6	10.9
Nd	0.9	0.1	0.7	1.0
Sm	0.3	0.1	0.2	0.4
Eu	0.9	0.2	0.6	1.1
Gd	0.5	0.1	0.4	0.5
Tb	0.4	0.1	0.4	0.7
Er	0.4	0.1	0.2	0.4
Yb	0.3	0.1	0.2	0.3

Notes: Data recalculated on an anhydrous basis. Values for oxides in weight percent; values for the trace elements, Cu, ppm; values for the rare-earth elements La to Yb in ppm. Number of samples given by n; St. Dev. = standard deviation; Min. = minimum value; Max. = maximum value.

On the AFM diagram (Figure 3-1), the basalts plot along on apparent iron-enrichment trend parallel to the gabbro-diabase trend but displaced to higher alkali contents.

Interpretation of Variation Diagrams

Effects of Alteration

In order to interpret major-element-variation trends some assessment of the effects of alteration and metamorphism on these elements is necessary. All samples show visible effects of alteration both in hand specimen and thin section.

The 'igneous spectrum' diagram of Hughes (1973) has been widely used to illustrate the mobility of the alkali elements during alteration. Unaltered igneous rocks plot within the igneous spectrum field while spilitic alteration will drive mafic rock compositions into a more soda-rich field above the igneous spectrum. It should be cautioned that a complimentary addition of both alkali elements in the proper proportions will move the composition towards more alkali-rich compositions within the igneous spectrum. The Pipestone Pond Complex mafic rocks are plotted on this diagram in Figure 3-4A. Three of the basalts plot within the igneous spectrum, which would seem to suggest that they have suffered little alkali mobility, although the remaining two plot above the igneous spectrum in the field of spilites. The gabbro and diabase samples, exhibiting very low alkali contents, generally plot within or near the igneous spectrum.

However, relationships shown on the K₂O-Na₂O diagram (Figure 3-4B) illustrate that the apparently unaltered condition of the basalts on the igneous spectrum diagram may be misleading. Most samples are notably enriched in alkali contents relative to normal oceanic and orogenic basalts. It is possible that these rocks were alkali rich when they were erupted; however, rocks with such alkalic original compositions should also show alkalic trends in other elements. In particular, they could be expected to be enriched in some or all of Rb, TiO₂, Ba or other incompatible elements. Inspection of the data shows that such is not the case, and it is therefore inferred that the basalts have suffered both soda and potash addition in proportions that have not moved the whole rock compositions out of the igneous spectrum field.

The diabase samples do not generally show Na₂O enrichment; only one sample is notably enriched compared to typical oceanic basalt compositions. However, all appear to be strongly K₂O depleted.

The possibility of CaO mobility can be tested by the Na₂O-CaO diagram of Miyashiro (1975). This diagram is useful only for rocks that represent liquid compositions. Basalts from the Pipestone Pond Complex are plotted on this diagram in Figure 3-4C. All show the effects of Na₂O addition and some are clearly CaO-depleted as well.

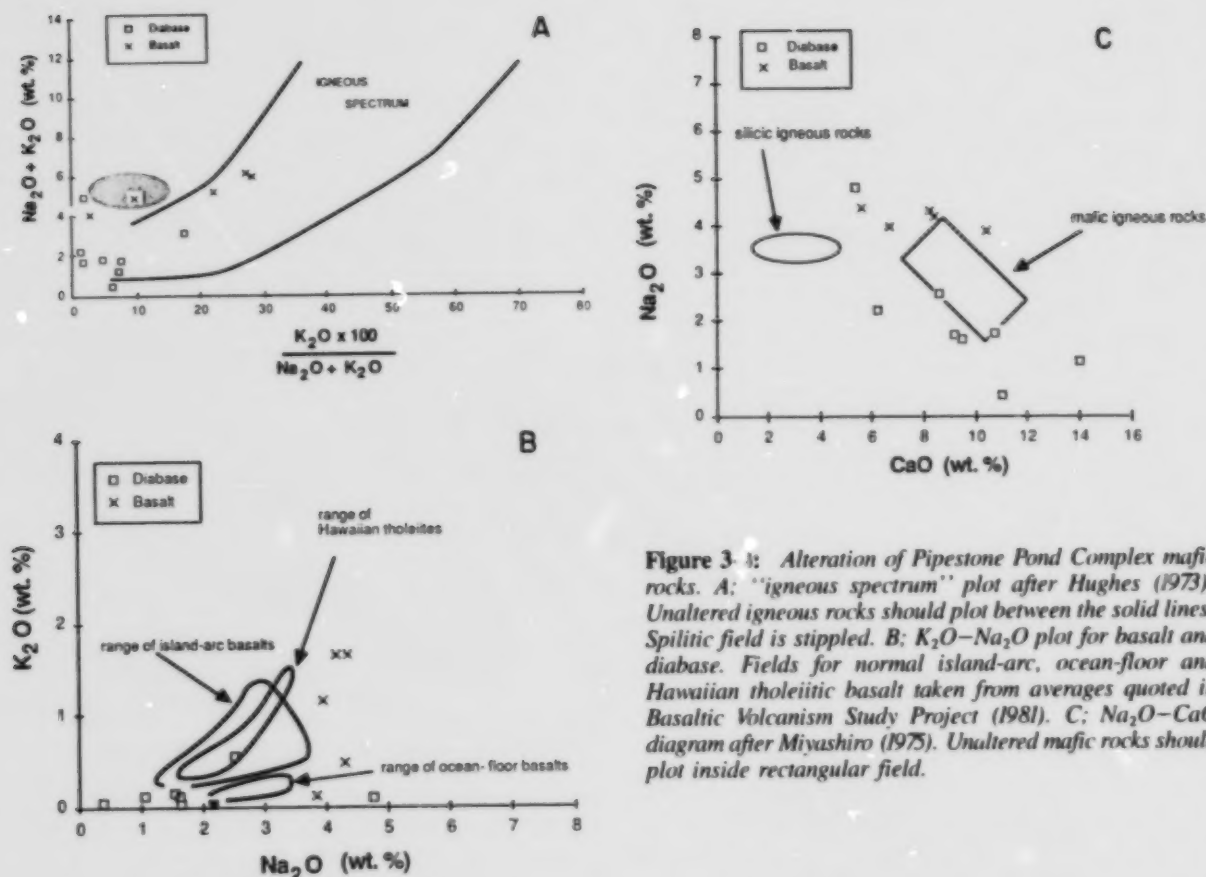


Figure 3-3: Alteration of Pipestone Pond Complex mafic rocks. A: "igneous spectrum" plot after Hughes (1973). Unaltered igneous rocks should plot between the solid lines. Spilitic field is stippled. B: K_2O – Na_2O plot for basalt and diabase. Fields for normal island-arc, ocean-floor and Hawaiian tholeiitic basalt taken from averages quoted in Basaltic Volcanism Study Project (1981). C: Na_2O – CaO diagram after Miyashiro (1975). Unaltered mafic rocks should plot inside rectangular field.

It is apparent from the preceding discussion that at least the basalts and diabases have suffered some major-element alteration. Diagrams involving CaO and the alkalis as well as low field-strength, trace elements such as Rb, Sr and Ba must therefore be interpreted with caution.

Magma Series of the Mafic Rocks

The separation of alkalic from subalkalic magma series in fresh basalts is commonly accomplished using the alkali-silica diagram of MacDonald and Katsura (1964). However, the alkali metasomatism documented in the previous section renders this diagram unreliable for the Pipestone Pond Complex basalts. Instead, reference is made to the TiO_2 – $\text{Zr/P}_2\text{O}_5$ diagram devised by Winchester and Floyd (1976) for use in altered rocks. On this diagram (Figure 3-5), both basalts and diabases plot in the nonalkalic field.

The nature of the magma series can be interpreted with reference to several diagrams. Keeping in mind the possible effects of element mobility, on the AFM diagram (Figure 3-1) the fractionation trend defined by the mafic intrusive rocks

of the complex is one of continuous iron enrichment, indicating that the fractionation series is tholeiitic in nature. The trend defined by the mafic intrusive rocks is very similar to the tholeiitic trend defined by the Annieopsquitch Complex of southwestern Newfoundland (Dunning, 1984, Dunning and Chorlton, 1985). However, on the SiO_2 – FeO^*/MgO and TiO_2 – FeO^*/MgO diagrams (Figures 3-6A,B), the diabase samples are widely scattered and do not define a clear continuous, fractionation trend. Although the least evolved (i.e., lowest FeO^*/MgO) samples appear to plot along a tholeiitic fractionation trend, it is not clear whether they are petrogenetically related to the more evolved samples.

All except one basalt sample plot within the tholeiitic field on the FeO^*/MgO – SiO_2 diagram and define a tholeiitic trend of increasing TiO_2 with differentiation on the TiO_2 – FeO^*/MgO diagram of Miyashiro (1975). The pillow lavas occupy a more restricted differentiation interval than the diabase dykes (FeO^*/MgO serving as a differentiation index) and plot along a steeper positive trend at higher TiO_2 concentrations. These basalts are unlikely to form part of a differentiation series that includes either type of diabase.

Tectonic Environment

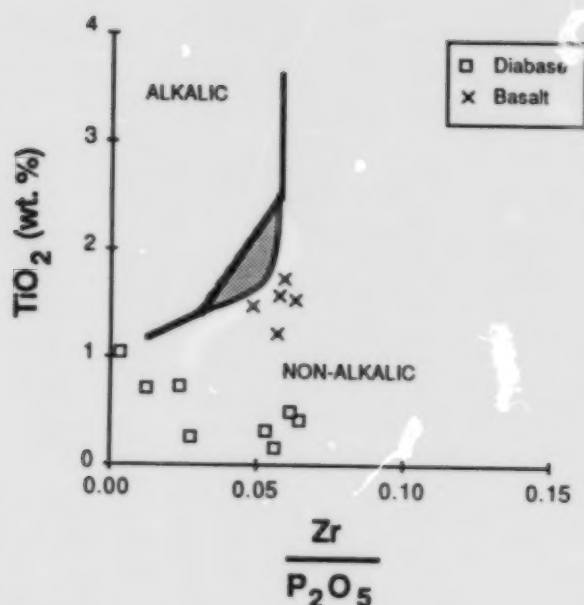


Figure 3-5: TiO_2 - $\text{Zr}/\text{P}_2\text{O}_5$ diagram after Winchester and Floyd (1976) illustrating the subalkaline nature of the Pipestone Pond Complex mafic rocks. Stippled field is transitional between alkalic and nonalkalic types.

Immobile trace elements have been used in various combinations by many workers to try and define the tectonic environments of different magma series. A review of the many diagrams that have been proposed is beyond the scope of this report. Many of these diagrams utilize elements for which the Pipestone Pond Complex has not been analyzed (e.g., Ta, Hf, Sc, Th). However, analyses are available for many incompatible (e.g., Zr, Ti, Y, Nb, REE) and compatible (e.g., Cr, Ni) elements, which are the least mobile during alteration. Many authors have emphasized that use of only one diagram or only one element pair can give a misleading interpretation in an inquiry into ancient tectonic environments because of the many variables affecting the concentration of all elements. Therefore, in the following section, a broad selection of element pairs and groups and a variety of diagrams will be presented in an attempt to constrain the tectonic environment represented by the Pipestone Pond Complex.

The Ti-Cr diagram was first used by Pearce and Cann (1973) to differentiate tholeiitic rocks of oceanic and island-arc affinity. On this diagram, rocks formed at a mid-ocean ridge would exhibit relatively elevated Ti contents compared to those in island-arc rocks at similar Cr contents. It can be seen from Figure 3-7A that basalts of the Pipestone Pond Complex plot close to the field boundary but mainly within the ocean-floor basalt field. The diabase samples have lower Ti contents and plot within the field of island-arc tholeiites.

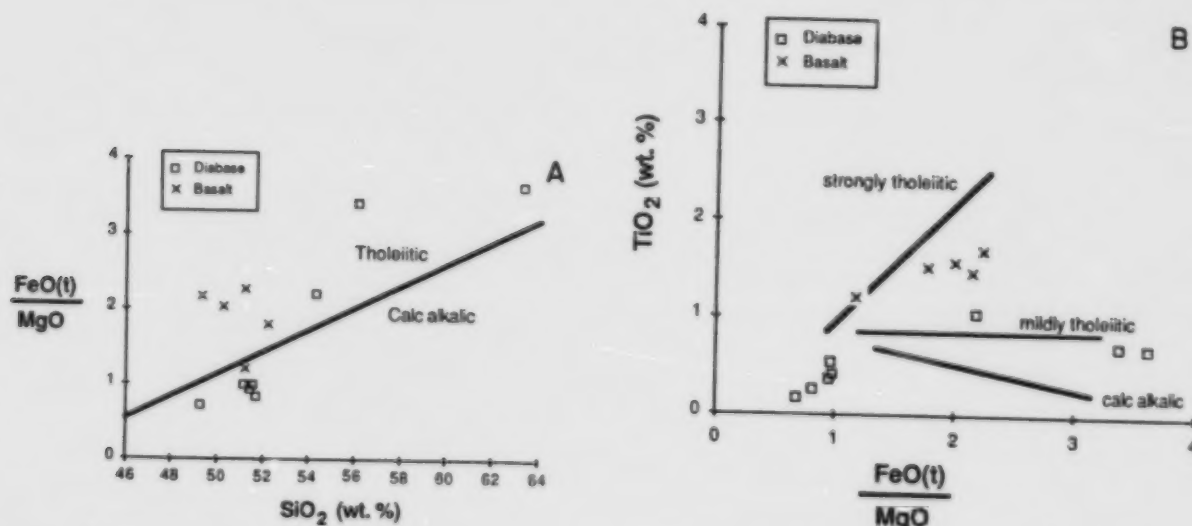


Figure 3-6: Magma series in the Pipestone Pond Complex basalt and diabase. A: FeO^*/MgO - SiO_2 diagram and B: TiO_2 - FeO^*/MgO diagram. Both confirm the tholeiitic nature of the magma series. In B, basalts have considerably higher TiO_2 contents at similar degrees of fractionation, suggesting that they are not from the same magma series as the mafic intrusive rocks.

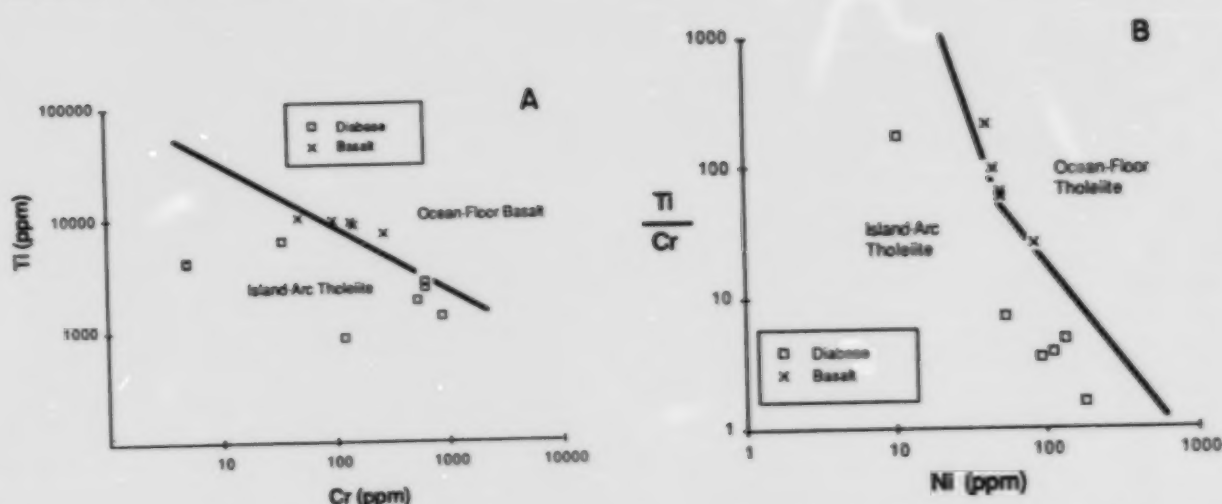


Figure 3-7: Tectonic environment discrimination of the Pipestone Pond Complex using Ti, Cr and Ni. A: Ti-Cr diagram of Pearce and Cann (1973); B: Ti/Cr-Ni diagram after Beccaluva *et al.* (1979). Basalts plot in the ocean-floor basalt field in both diagrams although very close to the field boundary with island-arc basalts. Dykes mainly plot within the island-arc basalt fields.

A similar diagram utilizing Ti/Cr and Ni was proposed by Beccaluva *et al.* (1979) (Figure 3-7B). They outlined fields of ocean-floor tholeiites and island-arc tholeiites, suggesting that the former were probably formed either at mid-ocean ridges or in marginal basins and the latter in immature island arcs. As in Figure 3-7A, basalts from the Pipestone Pond Complex plot very near the field boundary, but mainly within the ocean-floor basalt field, whereas the diabbases plot in the island-arc tholeiite field.

Many authors have made use of immobile trace element (e.g., Ti, Zr, Y, Ni, Cr) relationships to discriminate various oceanic basalts from those of calc-alkalic (island arc) affinity. Ti contents at a given Zr content should be lower in calc-alkalic basalts than in tholeiitic basalts. On the TiO_2 -Zr diagram of Pearce (1980), Pipestone Pond Complex basalts plot within the restricted field of mid-ocean ridge basalt (MORB) (Figure 3-8A). However, most diabase samples plot in a field of low Ti and Zr and plot outside the normal fields of oceanic volcanic rocks.

A similar result is evident on the Zr/Y-Zr diagram of Pearce and Norry (1979) and the Ti-Zr-Y diagram of Pearce and Cann (1973); (Figure 3-8B,C). The basalt samples plot in a trend that overlaps both the MORB- and the volcanic-arc basalt-fields. Most diabase samples, in contrast, have relatively Zr-depleted compositions and plot outside the island-arc fields, suggesting that they have been derived from sources that are even more depleted in incompatible elements than those typical in most island-arc environments.

Rare-earth-element (REE) concentrations for Pipestone Pond Complex basalts and diabase are shown in Figure 3-9. The diabase samples have distinctive REE patterns in which the light rare earths (LREE) are very depleted and the middle to heavy REE slightly less so. This contrasts sharply with

the slightly LREE-depleted, relatively smooth patterns of the Pipestone Pond Complex basalts. The latter have relative and absolute abundances similar to MORB and slight negative Eu anomalies. The different REE distributions in the diabase and basalt samples strongly indicates that the two are not petrogenetically related and probably represent different magma sources. This supports previous conclusions based on trace element diagrams.

Perfit *et al.* (1980) showed that some incompatible-element ratios (e.g., Ti/Y, Figure 3-10) may be characteristic of different types of oceanic basalts. The Pipestone Pond basalts have a Ti/Y ratio of approximately 200 similar to or slightly lower than typical MORB and also similar to that in modern island arcs. Diabase dykes appear to have slightly higher ratios, although Y concentrations in many of those samples are very low and of questionable precision.

In summary, the Pipestone Pond Complex basalts are part of a tholeiitic fractionation series. On most plots involving immobile trace elements, they show affinities to ocean floor basalt and/or MORB rather than island-arc lavas, but on some diagrams they plot very close to the field boundary. Immobile-trace-element abundances and REE distributions are similar to MORB.

The diabase dykes, most of which are believed to represent liquid compositions of the intrusive part of the complex, do not appear to be petrogenetically related to the basalts. They evidently represent a very depleted source; magmatic rocks representing such depleted sources are unlikely to represent a normal spreading-ridge environment, but the anomalous heat and/or water needed to melt such a source may be present in an island-arc environment or during the earliest phases of back-arc basin formation.

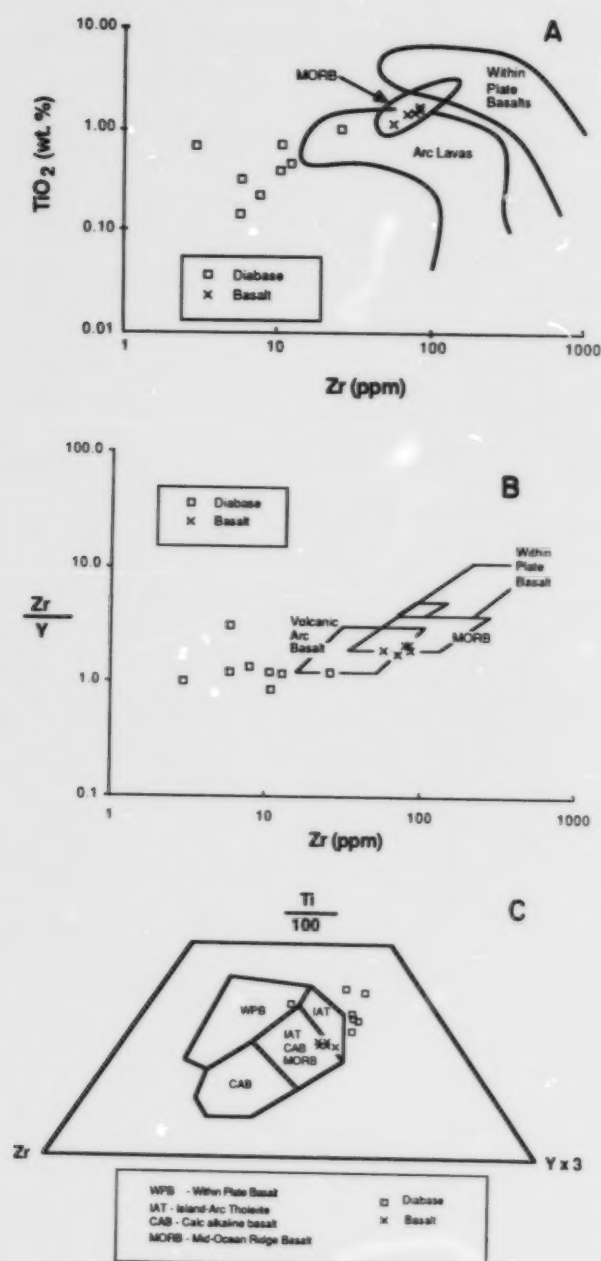


Figure 3-8: Discrimination of tectonic environment of the Pipestone Pond Complex using Ti, Zr, Y, Cr relationships. A: TiO_2 -Zr plot after Pearce (1980), B: Zr/Y -Zr plot after Pearce and Norry (1979), C: Ti-Zr-Y diagram after Pearce and Cann (1973). On all diagrams, basalts plot in or near the MORB fields, generally near the boundary with island-arc lavas while the dykes are depleted in incompatible elements, similar to island-arc lavas.

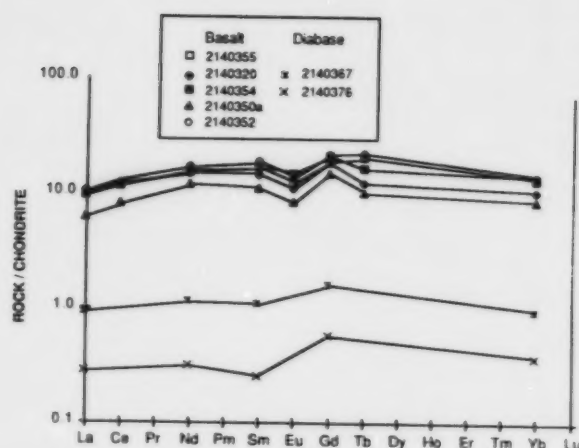


Figure 3-9: Rare-earth element plot for the Pipestone Pond Complex basalts and diabase dykes. Chondrite-normalization values are those of Taylor and Gorton (1977). Stipple field is range of MORB after Sun et al. (1979).

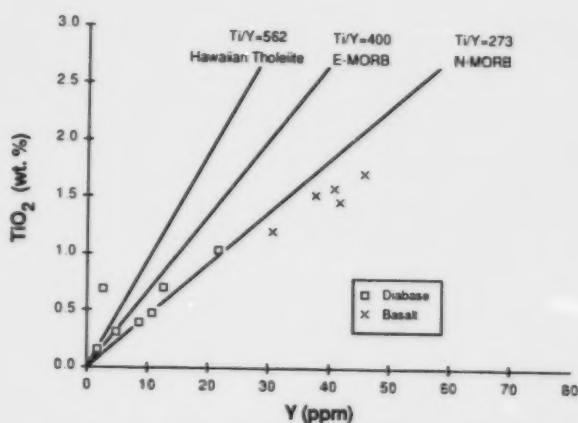


Figure 3-10: TiO_2 -Y plot for the Pipestone Pond Complex liquid compositions. Basalts are most similar to MORB.

Origin of the Plagiogranites

The origin of oceanic plagiogranites continues to be a matter of debate. Malpas (1979) showed that trondhjemites in the adjacent Bay of Islands and Little Port complexes in western Newfoundland have contrasting geochemistry, the latter being strongly depleted in the incompatible elements Ba, Y and Zr relative to the former. He explained this as being due to differences in genesis; the Little Port complex was interpreted as having formed through partial melting of amphibolite at the base of an island arc and the Bay of Islands trondhjemites through crystal fractionation of ophiolitic mafic magmas. Depletion of incompatible elements in the

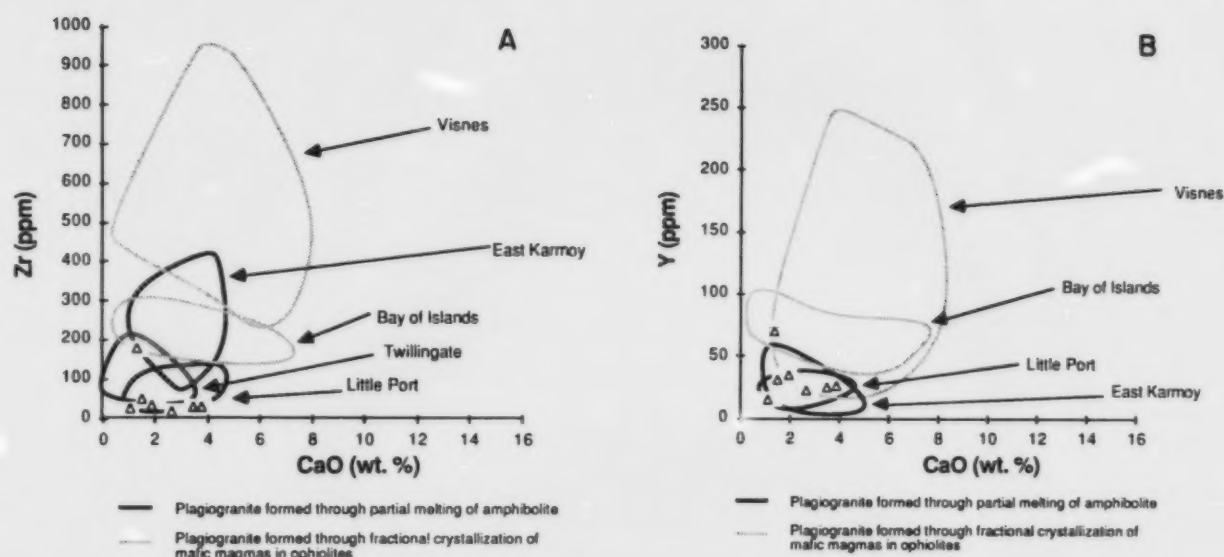


Figure 3-11: Incompatible-element relationships (A; Zr-CaO; B; Y-CaO) for Pipestone Pond Complex plagiogranites compared with those from the Bay of Islands and Little Pond Complexes (Malpas, 1979), the Twillingate trondhjemite (Payne and Strong, 1979) and the Visnes and East Karmoy complexes of the Karmoy ophiolite (Pederson and Malpas, 1985).

Twillingate trondhjemite similar to that in the Little Port Complex led Payne and Strong (1979) to suggest that this body also formed through partial melting of amphibolite at the base of an island arc. More recently, Pederson and Malpas (1984) have described a similar contrast in the chemistry of trondhjemites in the Karmoy ophiolite complex in Norway. There, a bimodal suite of plagiogranites can be related to two petrogenetic processes, fractional crystallization of mafic tholeiitic magmas in the ophiolite and partial melting of amphibolites in shear zones.

Incompatible-element (e.g., Zr, Y) abundances in plagiogranites of the Pipestone Pond Complex are similar to those in the Little Port complex, the Twillingate trondhjemite and the East Karmoy complex (Figure 3-11A,B) are considerably lower than would be expected through crystal fractionation of the spatially related gabbros. It is suggested that these plagiogranites have probably formed through partial melting of mafic rocks rather than through differentiation of the Pipestone Pond Complex gabbros. The mutually cross-cutting relationships between these rocks and other intrusive phases of the ophiolite necessitate that they were intruded at about the same time suggesting that mafic and felsic magmas were being generated contemporaneously in different source regions during formation of the Pipestone Pipe Complex.

COLD SPRING POND AND NORTH STEADY POND FORMATIONS

Mafic Volcanic Rocks

Nineteen samples of pillowed and massive basalt from within the outcrop area of the Cold Spring Pond and North

Steady Pond formations were analyzed during the present program. The analyses are presented in Appendix 2 and summarized in Table 3-7. Fifteen of the samples come from the newly defined North Salmon Dam Basalt. Of the remaining four, sample number 2140328 was collected from a mafic volcanic lens in the North Steady Pond Formation northwest of Pipestone Pond, 2140360 from a lens of mafic volcanic rocks associated with minor gabbro and plagiogranite in the Cold Spring Pond Formation southeast of Cold Spring Pond and 2140379 and 2140785 from the large ophiolitic sliver immediately north of Cold Spring Pond. These samples represent small, fault-bounded lenses of volcanic rock within the outcrop area of the North Steady Pond and Cold Spring Pond formations, but are not related to these formations stratigraphically. For convenience in the following discussion, they are termed the 'Sliver basalts'.

Most samples are basaltic (Table 3-7). Inspection of the data (Appendix 2) shows that the 'Sliver basalts' and the North Salmon Dam Basalt are clearly different chemically; the former generally contains less TiO_2 , Cr, Ni, Zr, Nb and P_2O_5 than the latter. In addition, sample 2140337 is clearly anomalous in most respects. The very high Cr and Ni contents of this sample suggest that it contains cumulative ferromagnesian minerals and does not represent liquid compositions. It is therefore excluded from the following discussion.

Alteration Effects

Alkali mobility in the Cold Spring Pond and North Steady Pond Formation basalts is suggested by the igneous spectrum diagram of Figure 3-12A. Most samples plot in the spilitic field above the igneous spectrum field boundary,

TABLE 3-7. Summary of geochemical analyses of basalts (North Salmon Dam and 'Silver Basalts') within the outcrop areas of the Cold Spring Pond and North Steady Pond Formations.

	North Salmon Dam Basalt (n=15)				'Sliver basalts' (n=4)			
	Mean	St.Dev.	Min.	Max.	Mean	St.Dev.	Min.	Max.
SiO ₂	48.8	2.6	43.9	51.6	52.0	23.0	50.4	55.2
Al ₂ O ₃	13.22	1.56	10.18	14.73	15.29	1.43	13.98	16.82
Fe ₂ O ₃	3.05	1.30	1.53	6.21	3.45	3.28	0.57	8.17
FeO	8.99	2.01	5.45	12.81	8.27	0.84	7.09	8.92
MgO	7.78	3.14	2.82	13.64	5.35	1.84	3.43	7.02
CaO	11.90	3.61	8.46	20.18	9.23	3.19	5.27	12.56
Na ₂ O	3.16	1.26	0.50	4.98	4.26	1.13	3.25	5.69
K ₂ O	0.39	0.29	0.09	1.23	0.54	0.51	0.08	1.06
TiO ₂	2.27	0.57	1.51	3.53	1.21	0.20	1.00	1.48
MnO	0.22	0.03	0.16	0.29	0.20	0.01	0.19	0.21
P ₂ O ₅	0.23	0.06	0.12	0.34	0.09	0.03	0.06	0.11
Cu	63	76	4	283	76	34	30	104
Zn	107	30	73	167	80	4	77	84
Ni	190	150	26	487	40	19	12	51
Rb	10	8	1	30	13	8	7	22
Sr	265	134	123	607	109	58	71	195
Y	31	10	21	60	31	7	22	36
Zr	134	37	83	225	51	22	25	73
Nb	18	7	9	39	5	1	3	6
Ga	13	3	10	20	10	3	7	13
Ba	98	56	42	243	161	146	22	360
V	267	116	106	505	398	78	314	474
Cr	318	208	36	618	107	31	61	132
La	9.3	0.9	7.9	9.8				
Ce	25.6	3.3	21.5	29.3				
Nd	17.1	2.6	14.7	20.1				
Sm	4.2	0.7	3.6	5.0				
Eu	1.3	0.2	1.0	1.5				
Gd	4.7	0.8	4.0	5.6				
Tb	0.7	0.1	0.6	0.8				
Yb	1.2	0.2	0.9	1.3				

Notes: Data have been recalculated on an anhydrous basis. Values for oxides in weight percent; values for the trace elements and rare-earth elements in ppm. Number of samples given by n; St. Dev. = standard deviation; Min. = minimum value; Max. = maximum value.

although four samples show evidence of possible K₂O addition and plot below the field. As illustrated on the K₂O–Na₂O diagram (Figure 3-12B), the principal alteration effect appears to have been Na₂O addition. However, the four anomalous samples on the igneous spectrum diagram do not show this Na₂O enrichment. They are variably depleted in Na₂O and one sample shows a prominent K₂O enrichment.

The effect of alteration on CaO is illustrated in Figure 3-12C. CaO does not appear to have been as affected as the alkalis in most samples, although one sample shows evidence of significant CaO depletion and four have had considerable CaO addition. Most of the North Salmon Dam basalts plot in or near the field of unaltered basalts.

Magma Series

Most Cold Spring Pond and North Steady Pond Formation basalts plot in the nonalkalic field on the TiO₂–Zr/P₂O₅ diagram (Figure 3-13). The minor overlap with the alkalic field is probably not significant as this diagram tends to overemphasize the alkalinity of rocks with transitional compositions (Goff, 1984). The North Salmon Dam basalts plot closer to the alkalic/subalkalic field boundary at slightly higher TiO₂ contents than do the Pipestone Pond Complex basalts. The three 'Sliver basalt' samples plot away from the field boundary in a similar position to the Pipestone Pond Complex basalt. On the AFM diagram (Figure 3-14), they plot along an iron enrichment trend similar to Hawaiian

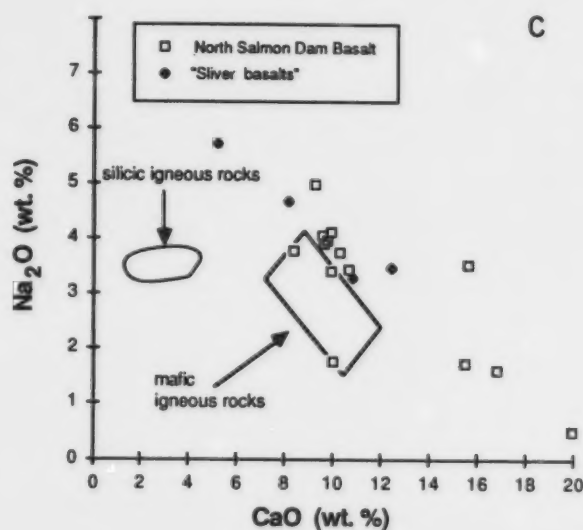
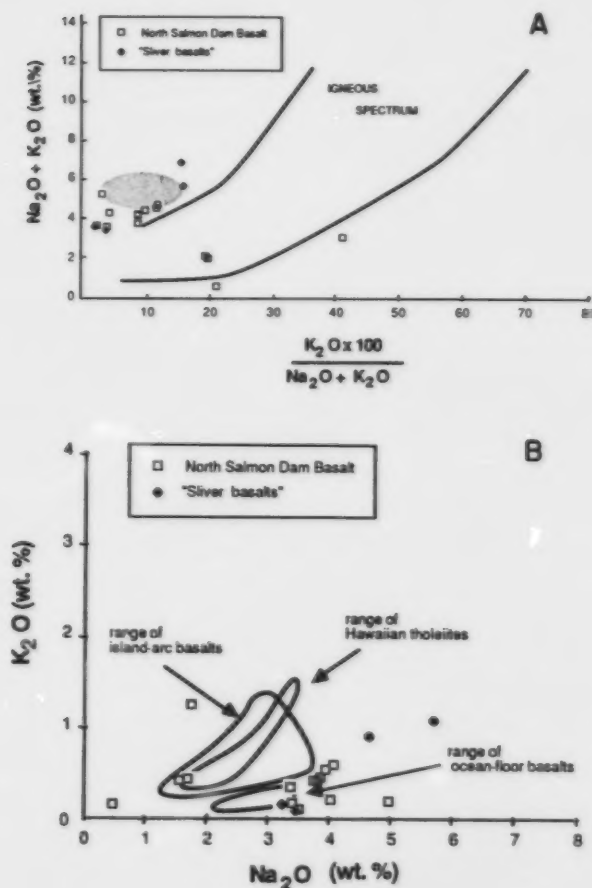


Figure 3-12: Alteration of mafic volcanic rocks within the outcrop areas of the Cold Spring Pond and North Steady Pond formations. A; "igneous spectrum" plot after Hughes (1973); B; K_2O – Na_2O plot; C; Na_2O – CaO plot after Miyashiro (1975). Field boundaries and sources as for Figure 3-14.

tholeiites. All samples plot within the tholeiitic field on the FeO^*/MgO – SiO_2 diagram (Figure 3-15A). On the TiO_2 – FeO^*/MgO diagram (Figure 3-15B), the North Salmon Dam basalts plot along a steep linear trend of rapidly increasing TiO_2 parallel to that of abyssal tholeiites (Miyashiro, 1975), confirming the tholeiitic nature of the fractionation series. The trend defined by the North Salmon Dam basalts is distinct from the slightly flatter, although still tholeiitic, trend defined by the Pipestone Pond Complex basalts. The two sequences clearly represent different fractionation series. The 'Sliver basalts' do not follow the North Salmon Dam Basalt trend but instead plot close to the field of the Pipestone Pond Complex basalts and mafic intrusive rocks. The similarity between these samples and the Pipestone Pond Complex will be further commented upon in the following section.

Tectonic Environment

All of the North Salmon Dam basalts plot well within the ocean-floor basalt field on the Ti – Cr diagram (Figure 3-16A), although the 'Sliver basalts' plot in and near the field defined by the Pipestone Pond Complex basalts and three of the four plot within the island-arc tholeiitic field. Likewise on the Ti/Cr – Ni diagram (Figure 3-16B), all the North

Salmon Dam basalts occupy the ocean-floor tholeiite field whereas the 'Sliver basalts' straddle the boundary between the ocean-floor tholeiite and island-arc tholeiite fields. In this respect the latter basalts are similar to mafic rocks of the Pipestone Pond Complex.

Most North Salmon Dam Basalt samples fall within the MORB field on the TiO_2 – Zr diagram (Figure 3-17A), although in a slightly more enriched part of the field than the Pipestone Pond basalts.

The geochemical distinction between the North Salmon Dam and Pipestone Pond basalts is more clearly displayed on the Zr/Y – Zr diagram (Figure 3-17B), in which the North Salmon Dam samples plot mainly in the lower part of the within-plate basalt (WPB) field. This emphasizes the elevated Zr/Y contents of this suite relative to the Pipestone Pond basalts. A similar separation of the two suites is evident on the Ti – Zr – Y triangular diagram (Figure 3-17C); the North Salmon Dam basalts plot within the WPB field, reflecting their relatively lower Y contents. On both diagrams, the 'Sliver basalts' plot in and near the island-arc fields and show affinities with Pipestone Pond Complex basalts and diabase dykes rather than the North Salmon Dam basalts.

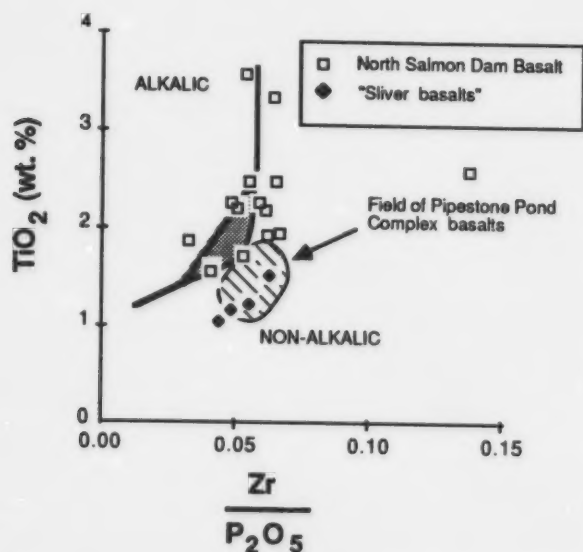


Figure 3-13: TiO_2 - $\text{Zr}/\text{P}_2\text{O}_5$ diagram after Winchester and Floyd (1976) illustrating subalkaline nature of North Steady Pond and Cold Spring Pond Formation mafic volcanic rocks. Stippled field is transitional between alkalic and nonalkalic types.

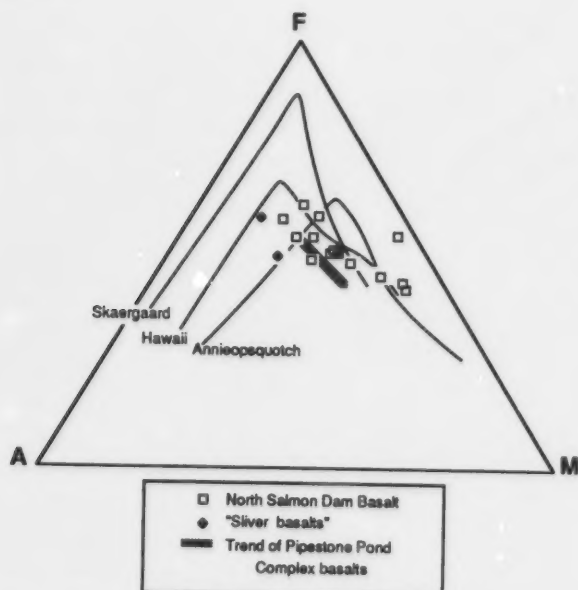


Figure 3-14: AFM diagram for basalts within the outcrops of the North Steady Pond and Cold Spring Pond formations.

REE distributions in the North Salmon Dam Formation basalts (Figure 3-18), in contrast to the Pipestone Pond Complex basalts, have a smooth negative slope with no Eu anomaly. These patterns suggest an affinity with more enriched magma sources, such as those of E-MORB (e.g.,

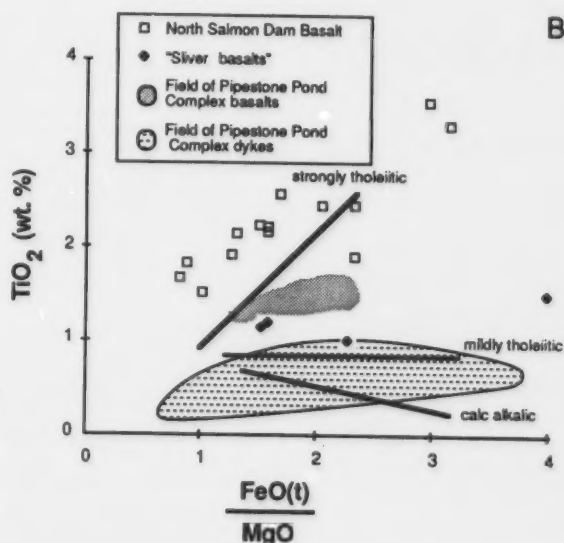
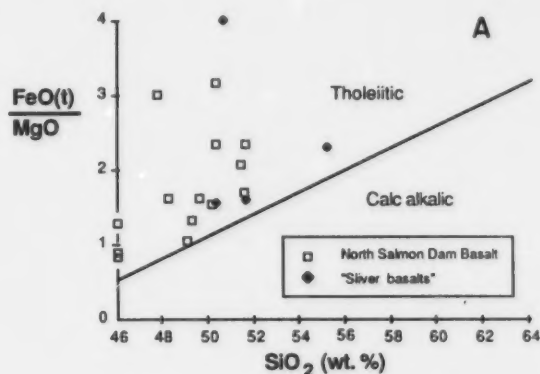


Figure 3-15: Definition of magma series for mafic volcanic rocks within the outcrop areas of the North Steady Pond and Cold Spring Pond formations. A: FeO^*/MgO - SiO_2 diagram and B: TiO_2 - FeO^*/MgO diagram. All are tholeiitic. In B, there is a clear distinction between the North Salmon Dam basalts and those representing outlying lenses within the two formations.

Sun, 1980) or oceanic island environments. They are consistent with trace-element distributions, which show a similar enrichment in the more incompatible elements. TiO_2 abundances are uniformly higher in the North Salmon Dam Basalt, and inspection of the data for the two suites shows that the North Salmon Dam Basalt also has uniformly higher P_2O_5 contents. These data, in particular the LREE enrichment, suggest a source area more enriched in incompatible elements.

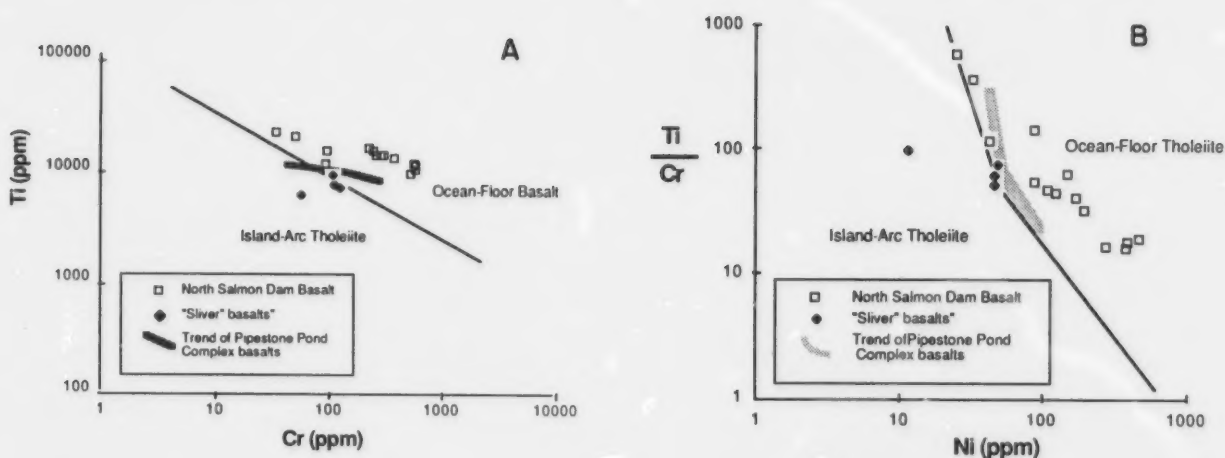


Figure 3-16: Discrimination of tectonic environment using Ti-Cr(A) and Ti/Cr-Ni(B) diagrams. Sources of diagrams as for Figure 3-8.

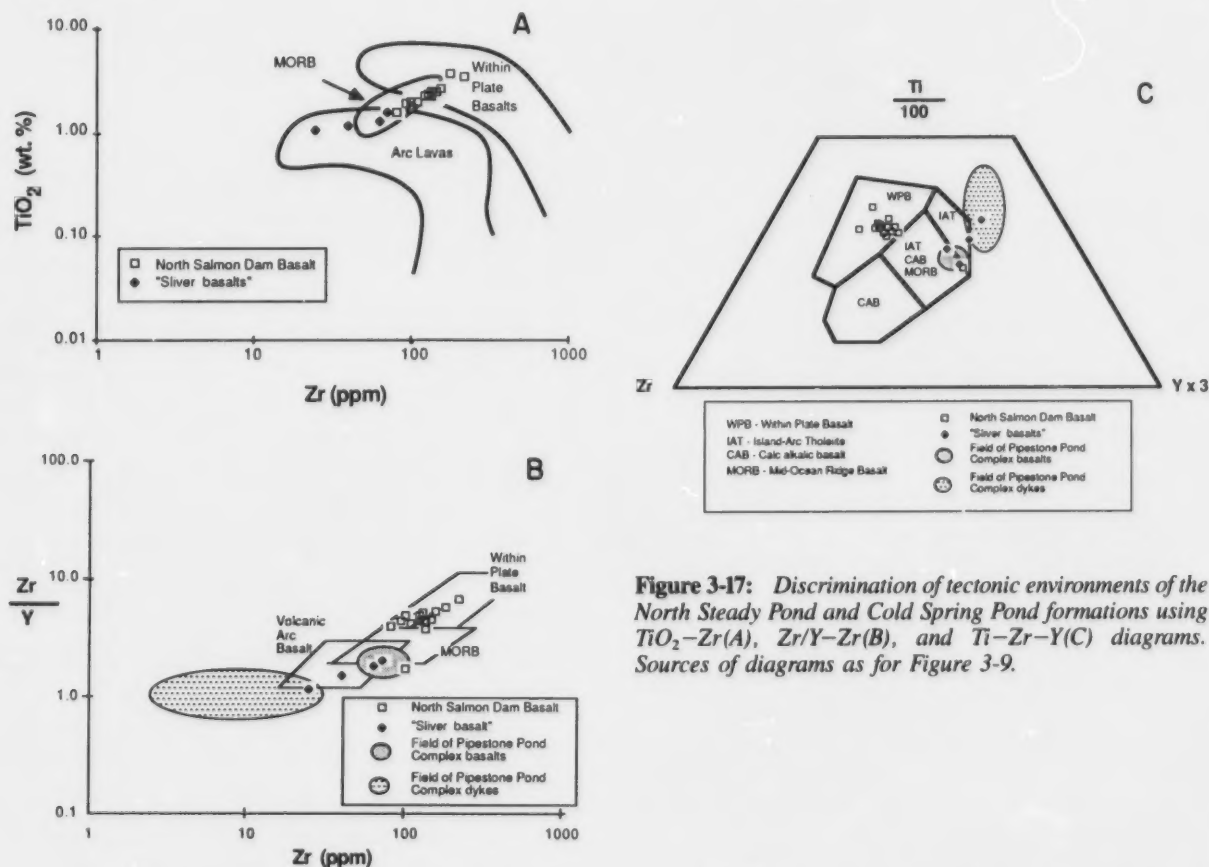


Figure 3-17: Discrimination of tectonic environments of the North Steady Pond and Cold Spring Pond formations using TiO₂-Zr(A), Zr/Y-Zr(B), and Ti-Zr-Y(C) diagrams. Sources of diagrams as for Figure 3-9.

The comparison of source areas for the Pipestone Pond Complex and North Salmon Dam basalts is further investigated in Figure 3-19. On a TiO₂-Y plot, the Cold Spring Pond basalts plot in a linear trend, which corresponds

to a Ti/Y ratio between 400 and 500. Perfit *et al.* (1980) have suggested that this ratio is typical of E-MORB. Significantly, the 'Sliver basalts' have Ti/Y ratios which are similar to N-MORB and also the Pipestone Pond Complex basalts.

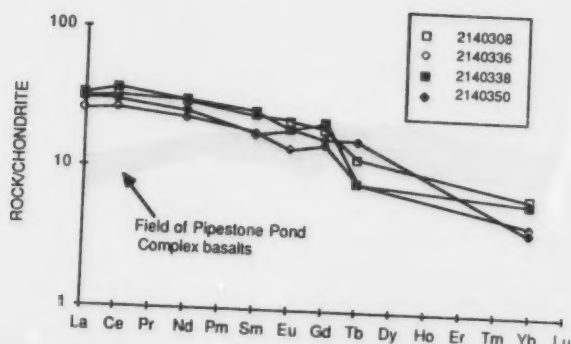


Figure 3-18: Rare-earth element plot of samples from the North Salmon Dam basalts. Chondrite normalization values are those of Taylor and Gorton (1977).

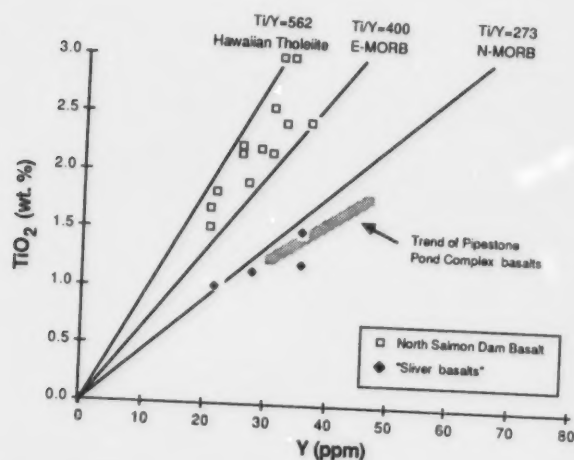


Figure 3-19: TiO_2 -Y plot for North Steady Pond and Cold Spring Pond basalts. North Salmon Dam basalts have ratios characteristic of E-MORB while those in outlying lenses are MORB-like, similar to those in the Pipestone Pond Complex basalts.

The chemical characteristics of the North Salmon Dam basalts strongly suggest that they did not form in an island-arc environment; in fact there is no evidence in any major- or trace-element relationships that would lead to this conclusion. Rather, the strongly tholeiitic nature of the basalts, coupled with the relatively high contents of incompatible elements (e.g., TiO_2 , Zr/Y, Ti/Y) strongly suggest that they are more similar to basalts generated in enriched segments of spreading axes (e.g., E-MORB) or as off-axis, within-plate seamounts in major ocean or marginal basins.

It is significant that of all the basalt within the outcrop area of the North Steady Pond and Cold Spring Pond formations, only the North Salmon Dam Basalt shows these characteristics. The analyzed outlying basalt lenses, are chemically similar to the Pipestone Pond Complex, and in some cases are closely associated with serpentinite, gabbro

and plagiogranite similar to lithologies within the Pipestone Pond Complex. The chemical characteristics coupled with the local association of rocks that are more clearly ophiolitic suggests that these basalt lenses occur in imbricate, fault-bounded slices rather than conformable lenses which were erupted during sedimentation of the enclosing rock units. Although not all basalts within the outcrop area of the North Steady Pond Formation have been analyzed, it may be that there are no volcanic rocks in the Baie d'Espoir Group in this area that are part of the depositional succession.

Felsic Volcanic Rocks

Felsic volcanic rocks occur within both the North Steady Pond Formation (subunit 8d) and the Cold Spring Pond Formation (subunit 10c). A representative suite of samples from each unit was analyzed to determine their relationship to each other and to the associated mafic volcanic rocks. A summary of analytical results is presented in Table 3-8.

Histograms of SiO_2 concentrations are presented in Figure 3-20. Felsic volcanic rocks of the Cold Spring Pond Formation are clearly more siliceous and are compositionally true rhyolites. North Steady Pond Formation felsic volcanics are mainly rhyodacitic to rhyolitic in composition. Those termed 'porphyritic dacite' in the field are generally rhyodacitic, whereas those termed 'rhyolite' are compositionally rhyolites.

Alteration Effects

Major elements in both units show considerable effects of mobility during alteration. On the igneous spectrum plot (Figure 3-21A), most samples plot to the right of the field of unaltered igneous rocks, indicating substantial relative sodium depletion. On the CaO - Na_2O diagram (Figure 3-21B), the samples plot in a field of relative Na_2O and CaO depletion relative to unaltered felsic volcanic rocks. The alteration has produced depletion trends in the analyzed felsic rocks similar to those in the Cold Spring Pond mafic volcanics (see previous section), suggesting similar processes of sub-seafloor alteration.

Petrogenesis

Field relationships, and lithological and petrographic considerations indicate significant differences between the felsic volcanic rocks in the two formations. These differences are also reflected in their geochemistry and suggest that the two are not part of a single fractionation trend. For example, in spite of the fact that some of the North Steady Pond Formation felsic volcanic rocks contain less SiO_2 than the Cold Spring Pond Formation rocks, the former are consistently enriched in incompatible elements (e.g., Ti, Zr, Y) relative to the Cold Spring Pond volcanics.

Because all analyzed mafic volcanic rocks within the outcrop area of the North Steady Pond Formation are thought to be faulted slivers of ophiolitic basalt (see previous section), no suitable parental magma for the North Steady Pond felsic

TABLE 3-8. Summary of geochemical analyses of felsic volcanic rocks in the Cold Spring Pond and North Steady Pond Formations.

	Cold Spring Pond Formation (n=6)				North Steady Pond Formation (n=8)			
	Mean	St.Dev.	Min.	Max.	Mean	St.Dev.	Min.	Max.
SiO ₂	73.6	1.2	72.3	75.6	71.5	6.3	66.3	85.6
Al ₂ O ₃	13.59	1.12	12.05	14.46	14.16	2.27	8.88	16.06
Fe ₂ O ₃	0.81	0.04	0.76	0.87	1.25	1.4	0.07	3.72
FeO	2.23	0.21	2.05	2.59	2.52	2.23	0	6.09
MgO	1.32	0.21	1.13	1.74	1.2	0.86	0.17	2.63
CaO	0.64	0.21	0.50	1.04	1.47	1.74	0.16	4.45
Na ₂ O	1.87	0.45	1.13	2.31	2.62	1.24	0.5	4.65
K ₂ O	5.17	0.46	4.71	5.93	4.69	3.28	0.12	8.88
TiO ₂	0.51	0.02	0.49	0.55	0.45	0.28	0.15	0.9
MnO	0.04	0.01	0.02	0.05	0.06	0.04	0.01	0.14
P ₂ O ₅	0.19	0.01	0.17	0.21	0.09	0.06	0.01	0.18
Cu	NA	NA	NA	NA	8	4	3	16
Zn	NA	NA	NA	NA	50	24	4	73
Ni	NA	NA	NA	NA	2	2	1	5
Rb	139	15	123	164	137	84	3	246
Sr	53	9	43	69	92	45	45	165
Y	45	6	37	55	61	11	46	77
Zr	177	13	160	192	234	85	87	323
Nb	15	2	13	17	12	5	1	16
Ga	19	2	15	22	14	6	1	24
Ba	687	115	583	868	724	457	77	1588
V	35	4	30	39	38	41	10	106
Cr	15	2	12	18	5	0	5	5

Notes: Data have been recalculated on an anhydrous basis. Values for oxides in weight percent; values for the metals Cu to Cr in ppm. Number of samples given by n; St. Dev. = Standard deviation; Min. = minimum value; Max. = maximum value; NA—not analysed.

volcanics can be identified. However, based on field association, the North Salmon Dam Basalt would be a potential parent to the associated felsic extrusive rocks in the Cold Spring Pond Formation. To test the possibility, the felsic volcanics are plotted on Zr-Y and TiO₂-Y diagrams (Figure 3-22A,B). In the former, Zr and Y contents are enriched over those in the basalt but the felsic rocks plot in a steep trend, which intersects the mafic volcanic trend near the most enriched basalt compositions. The difference in slope of the two trends and the fact that there is little difference in Zr content between the most evolved basalts and the least evolved felsic volcanics (representing a SiO₂ gap of more than 17%) indicates that the two are not petrogenetically related. There is also no obvious relationship between the two on the TiO₂-Y diagram as TiO₂ concentrations are much lower at only slightly higher Y concentrations, suggesting that the felsic volcanic rocks in the Cold Spring Pond Formation are not related by any simple fractionation scheme to the North Salmon Dam Basalt.

SUMMARY GEOCHEMICAL CHARACTERISTICS OF VOLCANIC ROCKS

The geochemical relationships between the various volcanic rocks in the field area lead to some conclusions regarding the nature of the volcanic activity:

1) The Pipestone Pond Complex basalts and the North Salmon Dam Formation basalts are chemically distinct. Although both display tholeiitic fractionation trends, the trend is stronger in the latter suite. The North Salmon Dam basalts are generally relatively enriched in incompatible elements, and their chemical composition is similar to E-MORB or within-plate, oceanic volcanism, e.g., hot-spot volcanism at a major spreading ridge or intraplate hot-spot volcanism. They show no evidence of processes related to subduction. The Pipestone Pond Complex basalts are chemically similar to N-MORB.

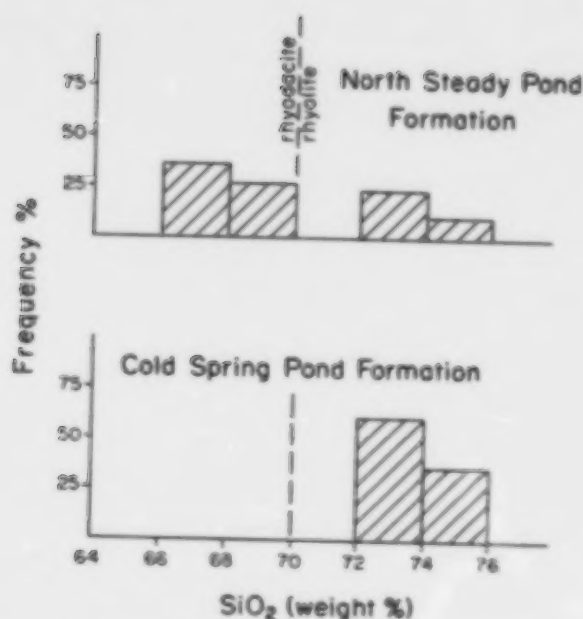
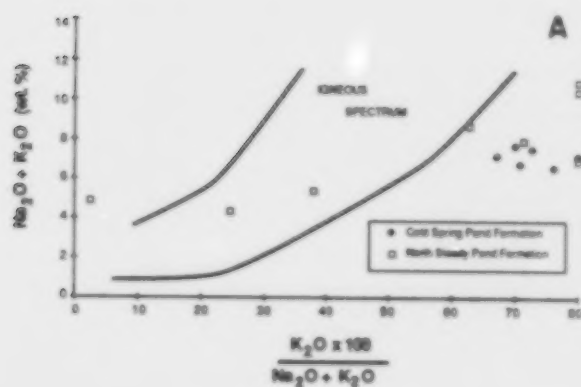


Figure 3-20: SiO_2 histograms for felsic volcanic rocks in the North Steady Pond and Cold Spring Pond formations. Cold Spring Pond volcanics are rhyolites, North Steady Pond volcanics include some rhyodacites.



2) Geochemical relationships between basalts and intrusive rocks of the Pipestone Pond Complex are not consistent with the derivation of the former from the latter by fractional crystallization. Some diabase dykes in the complex are very depleted in incompatible elements. They appear to represent melting of a very depleted source and are most likely to have been generated in a supra-subduction zone environment where dehydration of the subducting slab could result in hydrous partial melting of refractory sources in the overlying mantle wedge.

3) The lenses of mafic volcanic rock, which occur throughout the Cold Spring Pond and North Steady Pond formations west and southwest of the Pipestone Pond Complex (the 'Sliver basalts'), are chemically similar to the Pipestone Pond Complex basalts and/or diabase dykes and are chemically unlike the North Salmon Dam Basalt. They are, therefore, probably fault-bounded slices derived from the ophiolitic succession rather than stratigraphically conformable lenses within the sequence. Some of these basalts show close affinity with diabase dykes in the ophiolite, and may represent flows that are petrogenetically related to this magma type.

The geochemical relationships in these two sequences therefore support geologically based inferences that the Pipestone Pond Complex and the Cold Spring Pond Formation represent different Lower to Middle Ordovician geological environments that have been structurally juxtaposed. The differences documented above strongly

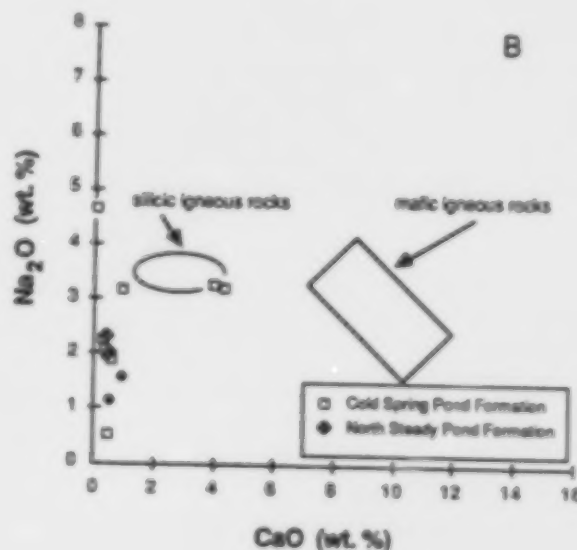


Figure 3-21: Alteration of felsic volcanic rocks. A: "igneous spectrum" plot after Hughes (1973) and B: Na_2O - CaO plot after Miyashiro (1975).

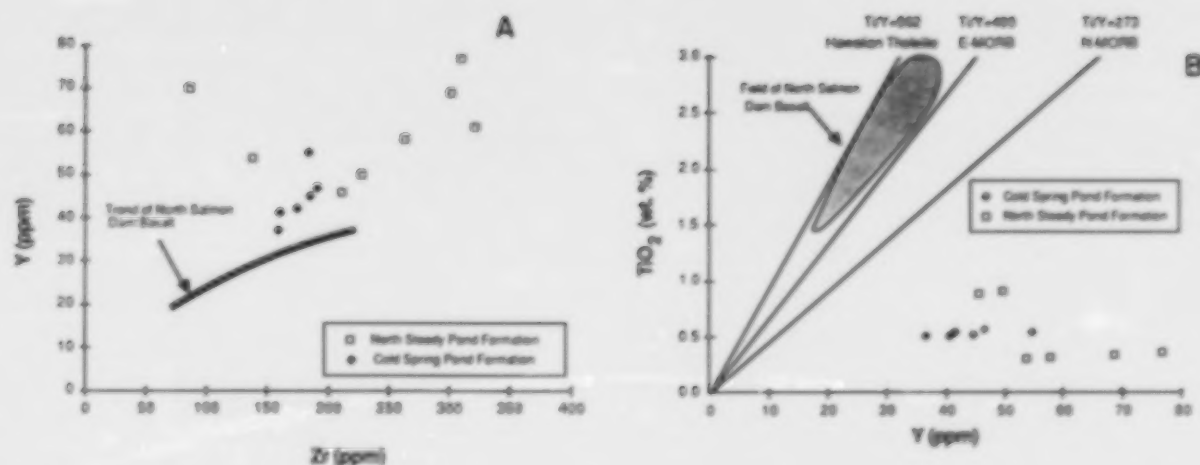


Figure 3-22: Y-Zr (A) and TiO₂-Y (B) plots for North Steady Pond and Cold Spring Pond Formation felsic volcanics.

suggest that the two sequences were generated under contrasting tectonic regimes; the Pipestone Pond Complex during the early stages of opening of a marginal basin, and the North Salmon Dam Basalt during oceanic intraplate volcanism. A reasonable hypothesis would be that the latter took place during a mature stage of development of the marginal basin; however, in the absence of radiometric age dates from the Cold Spring Pond Formation this remains speculative.

The relationship between felsic and mafic volcanics of the Cold Spring Pond Formation is uncertain. They display no obvious petrogenetic link; however, this would not be expected as there is seldom a direct fractionation relationship between mafic and felsic rocks in bimodal volcanic sequences.

CHAPTER 4

DEPOSITIONAL, METAMORPHIC AND STRUCTURAL HISTORY

DEPOSITIONAL HISTORY

The stratified rocks in each of the fault-bounded domains in the Pipestone Pond area are interpreted to be approximately coeval (Table 2.1). However, rocks in each of these blocks display significant differences in their depositional setting.

Spruce Brook Formation

Sandstones of the Spruce Brook Formation are compositionally and texturally mature, consisting almost entirely of quartz, with only accessory microcline, zircon, tourmaline and opaque minerals. There are no volcanic fragments and no evidence of a nearby volcanic source area. Colman-Sadd (1985a) and Colman-Sadd and Swinden (1984a) considered these rocks to have been deposited in a continental margin environment, suggesting that they had been eroded from sialic rocks of a cratonic area possibly located to the east. Colman-Sadd and Swinden (1984a) emphasized the similarity between these rocks and quartzitic psammities of the Gander Group to the east. They suggested that the two were deposited in similar environments and could perhaps be lateral equivalents.

Baie d'Espoir Group and Cold Spring Pond Formation

Siliciclastic rocks of the Salmon River Dam Formation in the Pipestone Pond area are similar to typical exposures of this formation in the nearby Twillick Brook and St. Alban's map areas (Colman-Sadd, 1980a, 1976 respectively). These rocks consist primarily of fine- to medium-grained sandstone, which display a to c divisions of the Bouma turbidite cycles. It is unclear from stratigraphic relations in the Pipestone Pond area if the Salmon River Dam Formation forms the base of the Baie d'Espoir Group as concluded by Colman-Sadd (1976, 1980a, 1980b) for other map areas.

Colman-Sadd (1980b) applied the submarine fan model of Walker (1978), to these formations, suggesting that they represent upper-fan to proximal mid-fan turbidites (North Steady Pond Formation) and distal, lower-fan turbidites (St. Joseph's Cove Formation). He suggested that the sediments of the North Steady Pond Formation were probably derived, at least in part, from the central Newfoundland island arc to the west. The Riches Island and Isle Galet formations outcrop along the eastern side of the Baie d'Espoir Group. Neither contains turbidites and the latter contains a considerable amount of felsic volcanic rock. Colman-Sadd (1980b) interpreted these formations to have been deposited at or near the continental margin at the eastern edge of Iapetus.

Stratified rocks of the North Steady Pond Formation contain both siliciclastic and volcanic rocks, which were interpreted by Colman-Sadd and Swinden (1982) and Colman-

Sadd (1985a) to form part of the central Newfoundland island arc. There are two principal siliciclastic lithofacies: 1) well-bedded psammitic and semipelitic turbidites locally interbedded with conglomerate and argillite; and 2) massive, unbedded, arkosic sandstones, pebbly sandstones and debris-flow conglomerates. The relationship between these two facies is exposed along the North Salmon River south of Pipestone Pond. Here, the basal sedimentary unit is coarse, polymictic, debris-flow conglomerate containing boulders of both ophiolitic- and island-arc-related lithologies. In its lower sections the conglomerate is unstratified but it gradually passes upward into thick Bouma cycles (a-b-c-e divisions) that thin and fine upward into c-e divisions where the North Salmon River turns sharply southward. In this instance, the debris flow-turbidite relationship forms an upward-fining succession. This sequence faces south and is faulted at its southernmost extent against rhyolites of the Huxter Pond member, which apparently form the top of a north-facing sequence of thin (c-e) turbidites overlain by a thick section of unstratified mass-flow sandstones and fine conglomerates. The stratigraphic relationship between the south- and north-facing parts of the North Steady Pond Formation is not known, although the lithologies in both sequences, particularly the conglomerates and fine turbidites, are very similar in aspect. In terms of the submarine fan models, they could be interpreted as representing a proximal upper-fan setting, close to a volcanic source area. In the adjacent Twillick Brook map area to the east, the North Steady Pond Formation is dominated by finer grained, more distal turbidites (Colman-Sadd, 1980a), suggesting that the provenance of this formation is probably to the west of the presently exposed units.

Stratified rocks of the Cold Spring Pond Formation include both felsic and mafic volcanic rocks as well as the dominant psammitic to pelitic turbidite successions. Turbidites in the Cold Spring Pond Formation are generally finer grained than in the North Steady Pond Formation, and east of Cold Spring Pond lack the characteristic debris-flow sandstones and conglomerates. Psammitic turbidites exposed along the North Salmon Dam access road north of the West Salmon River are similar in aspect to many fine grained units in the North Steady Pond Formation near Godaleich Lake. The turbidites apparently exhibit a gradual coarsening to the east, and in the area of Cold Spring Pond, include lenses of stratified, polymictic conglomerate, suggesting greater proximity to the volcanic source area somewhere to the west. The setting could be interpreted as mid fan to perhaps upper fan in the southwest.

However, recent models of volcanoclastic sedimentation in island-arc environments and back-arc environments (e.g., Karig and Moore, 1975; Carey and Sigurdsson, 1984; Klein, 1985) have emphasized that sedimentary patterns in these environments do not always conform to the idealized submarine fan models of, for example, Normark (1974) and

Walker (1978). Carey and Sigurdsson (1984) noted that island arc—marginal basin systems generally receive sediment from multiple sources along a (curvilinear) trend of volcanic islands, unlike typical submarine fan systems in which detritus emanates from a point source (submarine canyon) and is radially distributed. They further noted that enormous influxes of sediment from island-arc sources and random sediment dispersal patterns would characterize these sedimentary environments so that systematic facies transitions would be rare and difficult to interpret. They proposed a model for sedimentation in this environment, illustrated in Figure 4-1, which relates facies assemblages to tectonic conditions at various stages of back-arc basin development, identifies likely source areas and sediment types during different periods of back-arc development, and provides a conceptual framework for interpreting volcanoclastic sedimentary patterns in an environment where the submarine fan model may be inappropriate.

Their model (summarized below) considers sedimentary patterns in an island-arc—back-arc system to evolve through four evolutionary stages, each characterized by distinct facies assemblages, but not necessarily by consistent facies transitions.

Stage 1 (Figure 4-1A)—Initial rifting and formation of a marginal (inter-arc) basin: Basins formed by rifting tend to be narrow, steep-sided, fault-bounded troughs. Mass gravity flows, debris flows and turbidity currents inundate the basin floor. Large tectonic displacements characterize this environment and deep seated parts of the arc may be uplifted, denuded and exposed to sedimentary processes.

Stage 2 (Figure 4-1B)—Back-arc spreading and island-arc volcanism: Basin widening and subsidence results from continued back-arc spreading. Smoother-sided basins result from previous deposition of mass flow deposits. Development of a volcanoclastic apron adjacent to the island arc constructed principally from the products of renewed arc volcanism. Pelagic facies develop in the basin centre, which may prograde over the distal end of the volcanoclastic apron during lulls in volcanic activity.

Stage 3 (Figure 4-1C)—Basin maturity; Back-arc spreading wanes: Pelagic facies become more widespread in the basin as influx from the arc also wanes.

Stage 4 (Figure 4-1D)—Old age: A new cycle of marginal-basin formation begins. Influx of volcanoclastic debris from splitting of the arc may occur near basin margins but pelagic facies predominate elsewhere.

The depositional setting of volcanoclastic rocks in the Pipestone Pond area can be successfully interpreted in terms of the Carey and Sigurdsson (1984) approach. The model predicts that debris flow conglomerates containing detritus from both island-arc and probably ophiolitic sources are most likely to form in the steep-sided, fault-bounded basins during the earliest stages of back-arc basin opening. In this context,

the North Steady Pond Formation debris flow conglomerates assume some significance as they contain both island-arc and ophiolite derived detritus and are interpreted to depositionally overlie the Pipestone Pond Complex along the North Salmon River southwest of the Jamieson Hills. They may therefore record the earliest stages of sedimentation in a back-arc basin, the formation of which is recorded by the ophiolite. The model requires that there be an island arc which predates the Pipestone Pond Complex. Recently, evidence for pre-495 Ma, island-arc rocks is provided by a U-Pb (zircon) date of approximately 517 Ma for the Tally Pond volcanics (Dunning *et al.*, 1986).

According to the Cary—Sigurdsson model, the sources of ophiolitic detritus could either be newly formed back-arc basin crust or, more likely, somewhat older material uplifted by tectonic disturbance from the roots of the island arc. The coarse turbidites that overlie these conglomerates, as well as the mass flow arkoses and interbedded conglomerates that characterize the North Steady Pond Formation elsewhere in the area, are also predicted by the model to be present during Stage 1 of basin evolution. This lends credence to the applicability of the model in this particular case. The Salmon River Dam Formation, consisting predominantly of laminated quartzofeldspathic psammites, is less easily reconciled in the model because the composition of the detritus suggests a continental source. It appears to underlie the North Steady Pond Formation, of probable island-arc provenance, and may represent an early stage of sedimentation derived from a continent to the east. Fine grained turbidites of subunit 8a would perhaps represent more distal equivalents of the debris flow sediments and would be expected to interfinger with them as a result of changing rates of sediment influx from the island arc (Carey and Sigurdsson, 1984).

The sedimentary rocks in the Cold Spring Pond Formation are more commonly fine grained turbidites similar to those of subunit 8a, the most distal turbidites in the North Steady Pond Formation. Debris flow conglomerates and mass flow sediments are rare in the Cold Spring Pond Formation, which contains no recognizable ophiolitic detritus. In the Carey and Sigurdsson model, the Cold Spring Pond Formation might represent a more mature stage of back-arc basin development, perhaps forming part of the volcanoclastic apron generated during Stage 2 (Figure 4-1B). There is a considerable volume of black shale in this sequence as well, which may be analogous to the pelagic facies expected to develop in the basin centre as it widens and progrades with time over the distal ends of the volcanoclastic apron (e.g. Stage 3, Figure 4-1C).

Western Igneous—Metamorphic Domain

Sedimentary rocks in the structural block west of the Cold Spring Pond Formation (Unit II) are generally highly metamorphosed so their regional correlations are uncertain. The turbidites exposed in the narrow band northeast of Great Burnt Lake are very quartzose, and closely resemble sandstones of the Spruce Brook Formation. This is not

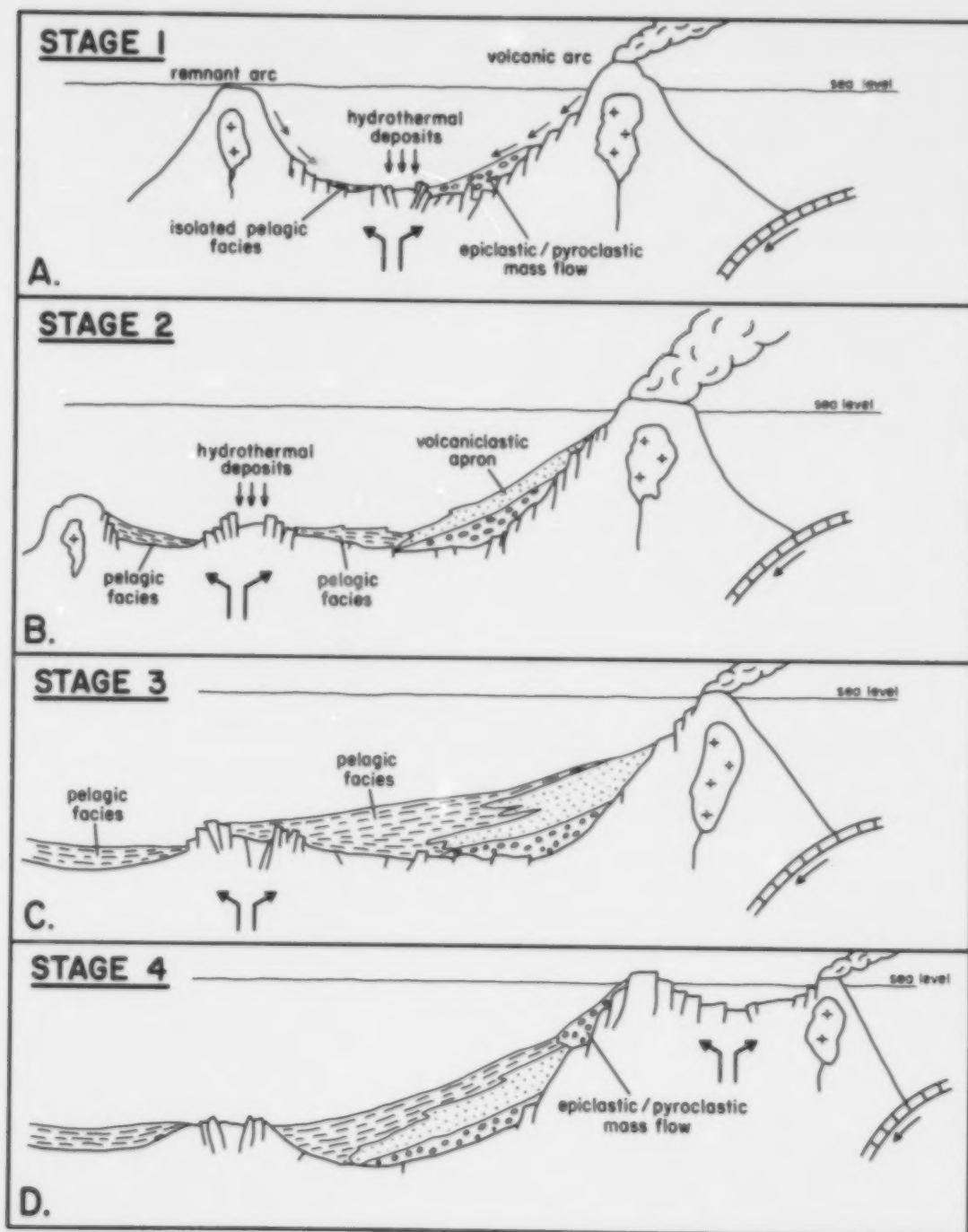


Figure 4-1: Model of the evolution of sedimentation (A to D) in island-arc-marginal-basin environments. After Carey and Sigurdsson (1984).

unreasonable as Colman-Sadd (1984, 1985b) and Dickson and Delaney (1984) have identified lithologies similar to the Spruce Brook Formation in map areas to the west and southwest of Great Burnt Lake. Elsewhere, psammities exposed in the area of the diversion channel linking Great Burnt Lake and Cold Spring Pond are purplish sandstones, similar to typical lithologies of the Salmon River Dam Formation. It may be that the metamorphic equivalents of lithologies from throughout the eastern margin of the mobile belt are tectonically interleaved in the western igneous-metamorphic domain. Detailed mapping of the various sedimentary units will be necessary to sort out their detailed associations and correlations.

METAMORPHISM

The intensity of peak metamorphic conditions varies widely across the Pipestone Pond area. In the western igneous-metamorphic domain, the peak condition was in the amphibolite facies, whereas the Cold Spring Pond Formation and the Baie d'Espoir Group were apparently metamorphosed only in the lower to middle greenschist facies. The Spruce Brook Formation exhibits a progressive metamorphic sequence ranging from lower greenschist to middle amphibolite facies. In all areas, the peak of metamorphism occurred during and shortly after the D_1 deformational event but prior to D_2 . Because these deformation episodes are considered to be correlative in all areas of the Pipestone Pond map area (see following section), it seems likely that the regional metamorphism is also correlative. The higher grades in the western domain probably indicate deeper burial during this regional metamorphic event.

The area east of the Pipestone Pond Complex forms part of the Mount Cormack Terrane, defined by Colman-Sadd and Swinden (1984a) as the area surrounded by the Pipestone Pond-Coy Pond-Great Bend ophiolite belt. The remarkable metamorphic progression displayed by the Spruce Brook Formation in this area is part of a more widespread metamorphic pattern in which metamorphic gradients increase regularly towards two centers of migmatization and granite intrusion. The metamorphic centers are exposed on the adjacent Burnt Hill map area and the metamorphic progression in this area has been described in some detail by Colman-Sadd (1985a). He suggested that the isograds shown on Figure 2-3 represent the following series of reactions:

biotite—this isograd represents the first appearance of biotite in quartz-chlorite-muscovite assemblages and probably formed through reactions involving changes in the composition of chlorite and/or muscovite.

andalusite—the appearance of andalusite coincides with the disappearance of chlorite and results from the reaction of chlorite, muscovite and quartz to produce andalusite, biotite and water.

sillimanite—sillimanite forms through the inversion of andalusite.

migmatite—represents melting near 650°C and 4 kbar.

The tectonic conditions to produce such a metamorphic progression have been discussed at some length by Colman-Sadd and Swinden (1984a) and Colman-Sadd (1985a) who, following Colman-Sadd and Russell (1982), attributed it to rising diapirs of anatectic melt. The melting was possibly related to the emplacement of the early thrust stacks (see following section) and the consequent burial of Spruce Brook-like lithologies to depths appropriate for partial melting to begin.

STRUCTURAL HISTORY

The structural history of the part of the Central Mobile Belt covered by this report has previously been discussed in detail by Colman-Sadd and Swinden (1984a) based in part on the field work reported here.

The earliest structural events recognized in the Pipestone Pond area are reverse faults, which juxtapose the Pipestone Pond Complex with the Spruce Brook Formation on the east and the Cold Spring Pond and North Steady Pond formations on the west. These faults dip moderately to steeply westward and are interpreted as eastward-directed thrusts (Colman-Sadd and Swinden, 1984a). The timing of this faulting is not tightly constrained. The maximum age is approximately Llanvirnian, based on the occurrence of fossils of this age in the Spruce Brook Formation which is affected by the faulting. The faulting is probably Late Silurian or earlier as the D_1 fabric is later than the faulting (see below) and affects rocks of this age. Colman-Sadd and Swinden (1984a) favoured a Middle to Late Silurian age for the faulting while noting that it could have been initiated much earlier and progressed through until this time, as proposed by Karlstrom *et al.* (1982) for the eastern Notre Dame Bay area to the north.

The principal regional deformation in the Pipestone Pond area is an east-west to northeast-southwest trending penetrative cleavage, which is axial planar to tight, upright, first generation isoclinal folds (F_1). This fabric is not deflected by the early thrust faults and probably postdates them. It is similar to and parallels the D_1 fabric in the Silurian Botwood Group sedimentary rocks and is therefore probably at least as old.

The two western domains in the Pipestone Pond area are cut by a second generation of reverse faults, which apparently postdate D_1 ; similar faults may be present in the Mount Cormack Domain but have not been recognized. One of these is the major north-trending fault that separates the Cold Spring Pond and the North Steady Pond formations; this fault truncates the D_1 fabric. A second major fault related to this set is the steep fault which juxtaposes the Cold Spring Pond Formation with the western igneous-metamorphic domain. This fault approximately parallels the D_1 fabric in this area but must be younger since the peak of metamorphism on either side of the fault postdated D_1 , which is apparently the same deformational event in both domains. It is unlikely that the upper greenschist-lower amphibolite facies rocks west

of the fault could have been juxtaposed with the lower greenschist facies rocks to the east during this metamorphic event; the fault must therefore have brought them into contact at some later time. There are many smaller faults of this nature throughout the area. West and south of the Pipestone Pond Complex, these faults are locally marked along their traces by ultramafic slivers, which serve as the principal means of recognizing them. Undoubtedly, other faults which do not have ultramafic rocks exposed along the trace are present but go unnoticed. Ultramafic rocks never appear on the traces of these later faults in the Mount Cormack Domain and this fact has been used by Colman-Sadd and Swinden (1984a) to support the concept that the ophiolitic rocks in this area are thrust over the Spruce Brook formation, and are themselves overthrust by the Cold Spring Pond Formation—Baie d'Espoir Group rocks. Ophiolitic rocks are believed to have been brought up along the traces of late faults only in the areas west and south of the Pipestone Pond Complex because only in these areas do they form the structural footwall to the exposed rocks; in the Mount Cormack Domain they are interpreted to have structurally overlain the Spruce Brook

Formation but to have been eroded later. Colman-Sadd and Swinden (1984a) noted that the northern and southern margins of the Pipestone Pond—Coy Pond ophiolite belt must have been modified by later faulting, because: 1) the full exposure of the ophiolite stratigraphy does not occur in these regions; and 2) because one of these faults, the Atlantic Lake—Miguels Lake fault, (Colman-Sadd and Russell, 1982), juxtaposes ophiolitic rocks with Silurian sedimentary rocks of the Wigwam Formation (Botwood Group); final movement on this fault must therefore be at least as late as Silurian.

The timing of the D_1 deformational event is not well constrained. In fact, there is no reason to believe that it is everywhere related to the same event. The variability in style and attitudes of related folds and planar fabrics suggests that second-generation structures developed mainly in response to local stress conditions and are not necessarily correlative across the map area. Colman-Sadd and Swinden (1984a) suggested that at least some of the second-generation folds may be related to the late steep faulting along the Atlantic Lake—Miguels Lake fault system.

CHAPTER 5

ECONOMIC GEOLOGY

GENERAL STATEMENT

The Pipestone Pond area is host to a number of significant mineral prospects of various types and has good potential for exploration for other deposit types that have not been reported to date from this area. Volcanogenic mineralization is widespread in the Cold Spring Pond Formation and there is further exploration potential for similar mineralization in the Pipestone Pond Complex, the Baie d'Espoir Group and the western igneous-metamorphic domain. Chromite, asbestos and magnesite occur in ultramafic rocks of the Pipestone Pond Complex. In addition, similar environments to those that host gold mineralization in southwestern Newfoundland may be present in the Baie d'Espoir Group, the Cold Spring Pond Formation and the western igneous-metamorphic domain.

Significant mineral occurrences are indicated on the geology map in the back pocket of this report and are described in more detail below.

PIPESTONE POND COMPLEX

The first reported concentrations of economic minerals in the Pipestone Pond area were the chromite occurrences east of Pipestone Pond hosted by sheared and altered peridotite of the Pipestone Pond Complex. These prospects have previously been examined and described by Snelgrove (1934), Fogwill (1964a) and Kean (1974). Chromite is widely disseminated in Unit 1_p, although it commonly forms less than 0.5 percent of the rocks (Plate 5-1). In rare instances it is concentrated in pods which are generally less than 2 m thick but which may reach up to 6 m in length. The largest concentration is found at the Chrome Pond prospect where a single massive chromite pod approximately 6 m long and averaging 0.5 to 1 m wide is hosted by sheared and serpentinized peridotite of Unit 1_p (Plate 5-2A,B). The showing occurs at the north end of a 100-m long outcrop ridge, which displays an unusually high chromite concentration along its complete length. Several smaller pods up to 1 m in length are to be found in the immediate vicinity of the main showing as are blebs of massive chromite and zones of heavily disseminated chromite. Analyses of chromite from this area by Fogwill (1964a) show that the material has a Cr:Fe ratio in the range 2.5 to 3.5; thus, if a deposit of mineable size could be outlined, there is every likelihood that the ore would produce a very high quality product.

Mapping throughout the ultramafic members of the ophiolite suggests that several other areas of slightly elevated chromite concentrations are present; none are known to be of economic proportions at the present time.

The sheared serpentinite of Unit 1_p and Unit 2_p also hosts minor occurrences of asbestos and magnesite. Asbestos generally occurs as minor concentrations of cross- and slip-



Plate 5-1: Disseminated chromite in Unit 1_p harzburgite locally forming thin, discontinuous bands.

fibre on widely spaced fracture surfaces (Plate 5-3). Although economically interesting quantities of asbestos have not been discovered in the Pipestone Pond area, a small deposit of good quality fibre has been discovered along strike to the northeast near Atlantic Lake in the Miguels Lake (2D/12) map area (Durocher, 1978). Exploration for magnesite has not been pursued in the Pipestone Pond Complex with the same intensity as elsewhere in the eastern Newfoundland ultrabasic rocks (see review in Colman-Sadd and Swinden, 1982). The present mapping suggests that development of magnesite is not as widespread in this area as, for example, in the Coy Pond and Great Bend areas. However, local concentrations are present, especially in Unit 1_p in the vicinity of Chrome Pond, and further exploration would seem to be warranted.

Recent interest in hydrothermally altered ultramafic rocks as gold targets (e.g., Buisson and Leblanc, 1986) suggests a further exploration potential for units 1_p, 2_p and 3_p of the Pipestone Pond Complex, as well as some of the larger outlying ophiolitic fault slices (e.g., Cold Spring Pond). Gold



Plate 5-2: *Chrome Pond prospect: A, shows a general view; B, a close-up with the chromite pod outlined.*



has been discovered elsewhere in eastern Newfoundland associated with altered ultramafic rocks of the Gander River Ultrabasic Belt, including showings at Great Bend (Zwicker and Strong, 1986), Jonathon's Pond (Blackwood, 1981) and Weirs Pond (O'Neill, 1987), lending further support to the exploration potential of the present areas.

The recognition of the Pipestone Pond Complex as an ophiolite raises the possibility that the pillow lavas, which underlie the western Jamieson Hills, and those in small ophiolitic slivers west of Pipestone Pond and north of Cold Spring Pond, may host volcanogenic sulphide deposits. Similar geological environments elsewhere in Newfoundland have been prolific copper producers (e.g., Whalesback, Little Bay, Tilt Cove, York Harbour; see reviews in Dean, 1977b; Swinden *et al.*, 1988), which are locally enriched in gold and also host gold showings (Swinden and Kean, 1984). No evidence of exhalative processes such as extensive alteration, exhalative iron- or manganese-rich horizons or sulphide concentrations were recognized during the present mapping. However, the basal part of the pillow lava unit would appear to be a good exploration target.

COLD SPRING POND FORMATION

There are two volcanogenic sulphide deposits in mafic volcanic rocks of the Cold Spring Pond Formation (the North Salmon Dam Basalt). The larger of these, the Great Burnt Lake deposit, is located on the southeastern side of a prominent hill approximately 2 km northeast of the North Salmon Dam; the smaller deposit, the South Pond deposit, is located approximately 10 km to the north of the Great Burnt Lake deposit and 3 km southwest of South Pond. Both presently lie within the boundaries of the South Pond mining lease (Figure 1-1), previously held by Abitibi-Price Incorporated and Asarco Limited and, since 1985, by BP-Selco Limited. The sulphide zones are not exposed, however, both have been extensively drilled. The following descriptions are based on examination of drill core from several holes which cut the deposits.

The Great Burnt Lake deposit is a roughly tabular, conformable body of massive pyrrhotite-chalcopyrite-pyrite mineralization. The geological sequence in the vicinity of the deposit consists of mafic flows and tuffs with interbedded



Figure 5-3: Widely spaced fractures (perpendicular to knife) containing asbestos cross fibre in Unit 1_p harzburgite.

siltstone and greywacke and fine grained clastic sediments, which are locally graphitic. The deposit ranges up to 14 m in thickness, strikes approximately 030°, with steep dips to the southeast, and plunges gently to the southwest. The main mineralized zone carries sulphide concentrations ranging from less than 40 percent to 90 percent. Pyrrhotite is the dominant sulphide mineral present and generally occurs as massive beds. Chalcopyrite is distributed throughout the mineralized zone, forming large blebs within the massive pyrrhotite, as well as discrete strata of massive mineralization up to 6 cm thick (Plate 5-4). To the west, the sulphide zone is in sharp contact with quartzose greywacke and siltstone, which shows little evidence of alteration and/or mineralization. To the east, the deposit is in contact with pillow lavas, although a mafic volcanic breccia up to 1.5 m thick locally occurs between the volcanic rocks and the massive mineralization. The volcanic sequence to the east of the massive mineralization exhibits widespread although not particularly intense, alteration within 30 to 40 m of the massive zone. Black chloritization (and associated stringers of pyrite, pyrrhotite and chalcopyrite) and local silicification are the principal alteration effects. Similar alteration and

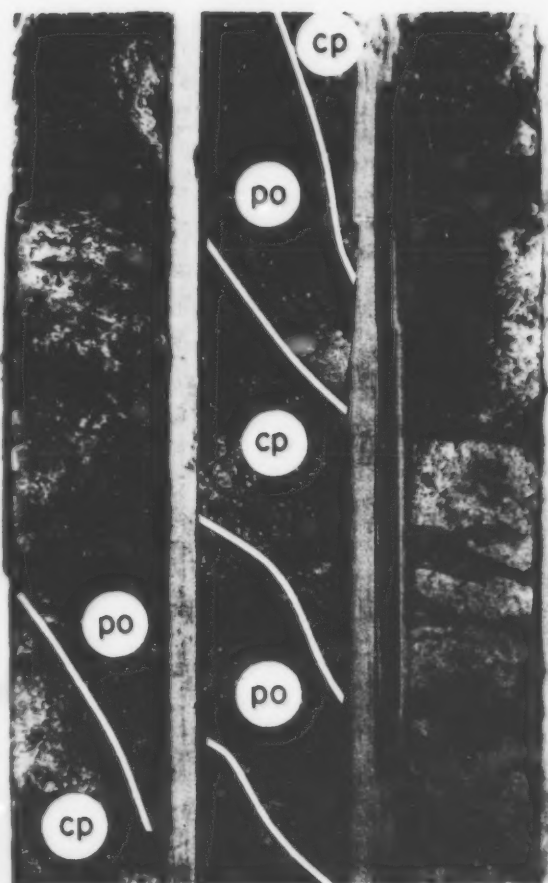


Figure 5-4: Bedded sulphides in drill core from the Great Burnt Lake deposit. Alternating beds of chalcopyrite (cp) and pyrrhotite (po) are outlined.

sparse, erratic mineralization are prevalent in an apparently separate diffuse zone up to 335 m long and 75 m thick north and east of the massive zone. The geological situation is suggestive of syngenetic sulphide mineralization on the sea floor at the top of a pillow lava succession and followed by a period of clastic sedimentation. The extensive alteration and sparse mineralization in rocks on the eastern side of the massive sulphide body may record the passage of hydrothermal fluids related to the mineralizing event and therefore suggest that the deposit is slightly overturned and facing to the northwest.

The Great Burnt Lake deposit is subeconomic at present. Recent estimates suggest that it contains on the order of 750,000 tonnes of 2-3 percent copper.

The South Pond deposit lies approximately 10 km north of the Great Burnt Lake deposit and occupies approximately the same stratigraphic interval. Sulphides are stratigraphically

more widespread in this area and in some cases are present virtually throughout some drill cores. Pyrrhotite commonly comprises 0.5 to 1 percent in tuffaceous rocks and up to 20 percent over short sections. Chalcopyrite disseminations and blebs locally accompany the pyrrhotite; a particularly chalcopyrite-rich zone has been outlined, which constitutes the South Pond deposit. This zone consists of up to 15 m of mafic tuff, containing up to 30 percent sulphides and consisting of pyrrhotite, which forms anastomosing layers semimassive zones and disseminations, and chalcopyrite blebs up to 1 cm in diameter. The boundaries of the zone are gradational, reflecting both a decrease in total sulphide content and an attendant disappearance of chalcopyrite.

The South Pond deposit is considerably smaller than the Great Burnt Lake prospect. It is estimated to contain about 300,000 tonnes grading approximately 1 percent copper.

Although neither of these deposits is of economic size or grade, they are important indicators that exhalative processes accompanied this volcanism. Furthermore, mineralization is not restricted to the two known deposits in the volcanic belt; minor amounts of pyrrhotite and chalcopyrite have been noted in drill core between the two deposits. Excavation during construction of the North Salmon Dam exposed an outcrop of intensely chloritized basalt on the south shore of the North Salmon River, which contained minor disseminated chalcopyrite. This occurrence, along strike with the Great Burnt Lake prospect 2 km to the north, was covered over by subsequent construction activity. This showing was apparently not of economic dimensions or grade. However, it indicates that sulphide mineralization is widespread at this stratigraphic interval and it would appear that the whole strike extension of the volcanic belt should be considered as favourable prospecting territory.

BAIE D'ESPOIR GROUP

The North Steady Pond Formation of the Baie d'Espoir Group is considered to have good potential for exploration for volcanogenic base- and precious-metal deposits. The principal target lithology is felsic volcanic rocks of subunit 8d, primarily dacitic to rhyodacitic flows and ash flows which locally contain coarse breccias and agglomerates suggestive of nearby explosive volcanism. The felsic volcanic rocks locally contain small (several square metres) alteration patches characterized by pyrite and sericite. Although none of these were found to contain base- or precious-metals in the present area, a single sample from the Burnt Pond area immediately to the east assayed 10 g/t Ag (Colman-Sadd, 1985a). Some exposures of the volcanic rocks show evidence of being deposited in subaerial conditions, but the volcanics are part of a succession that include turbidites. It seems likely, therefore, that they formed volcanic islands and, as such, they present a highly favourable environment for massive sulphide deposits resulting from submarine fumarolic activity.

Volcanogenic sulphide deposition is known from similar environments elsewhere in southern Newfoundland. The Strickland prospect and several associated occurrences in the La Poile Bay area of southwestern Newfoundland (Cooper,

1954; Stackhouse, 1976; Swinden, 1981a) occur in a volcano-sedimentary succession, the volcanic component of which is dominated by felsic rocks. Likewise, the Barasway de Cerf prospect in southern Bay d'Espoir (Jewell, 1939; Swinden, 1980a, 1982), although in a relatively basinal environment, is closely associated with pulses of felsic volcanism. Massive sulphide mineralization associated with felsic volcanism in southern Newfoundland is dominantly Pb-Zn-Ag-rich, containing lesser amounts of pyrite and copper. Swinden and Thorpe (1984) have suggested from geological and lead-isotopic evidence that this reflects the influence of continentally derived material in the source areas for the metals. A similar source region is suggested for Unit 8d by its similarity to the other Hermitage Flexure volcanic sequences and by analogy, it might be expected that mineralization associated with this member would have a similar mineralogical composition; this could have implications for the exploration methods most likely to be successful in this terrain.

A further potential for these rocks is suggested by the recent discovery of gold mineralization in the Cinq Cerf Brook area of southwestern Newfoundland. The mineralization has some characteristics of a deformed epithermal alteration zone related to felsic volcanism and subvolcanic intrusion (Swinden, 1984; McKenzie, 1986). This discovery has sparked considerable exploration interest in similar lithologies of the Isle Galet Formation in the Baie d'Espoir Group and elsewhere around the Hermitage Flexure. Unit 8d is considered to represent a very similar geological environment, and in the absence of more specific notions on the controls on gold mineralization in this environment, it must be considered as a high-potential target for gold exploration.

WESTERN IGNEOUS-METAMORPHIC DOMAIN

The economic potential of the western igneous-metamorphic domain to the west of the Cold Spring Pond Formation is not well defined. This area was not investigated in detail during the present program, and a detailed economic assessment is therefore beyond the scope of this report. However, a few brief comments on economic potential are given below.

The metamorphic rocks in this area appear to be dominantly metasedimentary and bear no obvious economic potential. However, northwest of Cold Spring Pond, they include some buff to white, very silicic rocks which appear to be metamorphosed felsic volcanics similar to those which form part of the Isle Galet Formation to the east (e.g., Colman-Sadd and Swinden, 1982). These were only seen in a few outcrops and recent mapping by Colman-Sadd (1984, 1985b) suggests that they are not widespread. Exploration including diamond drilling for volcanogenic sulphide mineralization was carried out in this area by Riocanex (Bucknell and MacKenzie, 1977a,b) with negative results. The environment, however, seems broadly similar to the North Steady Pond and Isle Galet Formations and a similar base- and precious-metal-exploration potential may exist.

Granitoid rocks in this domain are predominantly highly deformed biotite ($\pm$ hornblende) varieties that do not display any obvious economic potential.

SPRUCE BROOK FORMATION

There are no known mineral occurrences in the Spruce Brook Formation in the present area and mapping during this

project has not demonstrated any obvious exploration potential. Zwicker and Strong (1986) have recently discovered auriferous arsenopyrite- and pyrite-bearing slate having anomalous gold values in equivalent sedimentary strata near the margins of the Great Bend Complex. Although the relationship of the mineralization to stratigraphic or structural setting or to the presence of the nearby ophiolite is not clear, it suggests a potential for similar gold-bearing rocks in the Pipestone Pond area.

CHAPTER 6

CONCLUSIONS

A number of geological relationships in the Pipestone Pond area have been clarified as a result of this study. Some of these are of considerable importance with respect to regional geological relationships in the Newfoundland Central Mobile Belt. Some of the significant results of this work are summarized below with brief mention of their possible regional significance.

1) The ophiolitic nature of the Pipestone Pond Complex has been documented and correlations established with the Coy Pond and Great Bend complexes to the east. The documentation of the structural contact between the ophiolite and the underlying Spruce Brook Formation has led to new interpretations of the structural history of the eastern part of the Mobile Belt, including likely post-Taconian, thin-skinned thrust tectonics (Colman-Sadd and Swinden, 1984a).

2) Volcanic and sedimentary rocks east of Great Burnt Lake, part of what was previously termed the 'Great Burnt Lake volcanic belt', have been mapped and separated from both the Pipestone Pond Complex and the Baie d'Espoir Group.

3) A depositional contact between the ophiolitic rocks and the Baie d'Espoir Group is proposed. Basal debris-flow conglomerates of the Baie d'Espoir Group contain detritus derived both from ophiolitic and island-arc sources.

4) The remarkable metamorphic gradients documented by Colman-Sadd (1981) in the Spruce Brook Formation in the Burnt Hill area have been identified in the present area. Metamorphic isograds increase from west to east in this area and the maximum metamorphic grade is marked by the andalusite isograd.

5) The eastern margin of the western igneous-metamorphic domain has been mapped and shown to be marked by an abrupt transition eastwards from lower amphibolite to greenschist facies. The tectonic nature of this contact is confirmed by slivers of ophiolitic material along the fault trace.

6) Mafic and ultramafic lenses within the Cold Spring Pond Formation are interpreted as fault-bounded slivers brought up along steep reverse faults. This implies that ophiolitic material underlies the central mobile belt at least as far as the western limit of mapping.

7) Geochemical relationships within the Pipestone Pond Complex, Cold Spring Pond and North Steady Pond formations have been documented. Among the more important conclusions are:

a) Geochemical data suggest that the Pipestone Pond ophiolite may have formed in a supra-subduction zone environment. The various intrusive and extrusive parts of the ophiolite are not obviously linked petrogenetically.

b) Basalts in the Cold Spring Pond Formation, distinct from those in the Pipestone Pond Complex, are interpreted to represent volcanism in an oceanic within-plate setting or an anomalously enriched ridge segment, possibly in a back-arc basin.

c) Isolated lenses of basalt in the Cold Spring Pond and North Steady Pond formations are related to the Pipestone Pond Complex and not part of the depositional sequence recorded by the surrounding rocks.

8) Application of recent models of sedimentation in back-arc basin environments suggests that sediments of the North Steady Pond and perhaps the Salmon River Dam formations formed during the initial stages of back-arc basin opening. Those in the Cold Spring Pond Formation appear to be more typical of mature back-arc basins, suggesting that, if they formed in the same back-arc basin, the Cold Spring Pond Formation may be slightly younger. The sedimentary models are in accord with interpretations of tectonic environments based on geochemistry of the associated volcanic rocks.

9) Volcanogenic mineralization in the Cold Spring Pond Formation appears to have taken place in a back-arc environment associated with volcanism removed from the subduction zone. This probably accounts for the lack of lead and zinc in the deposits.

10) Environments similar to the gold-bearing environments in southern Newfoundland are present in the North Steady Pond Formation and perhaps in the Cold Spring Pond Formation and the western igneous-metamorphic domain. Good exploration potential is indicated in these units.

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APPENDIX 1

SAMPLING AND ANALYTICAL METHODS

Samples for geochemical analysis were collected in the field with an 8-pound sledge hammer. Approximately 2 kg of rock were collected from the fresh interiors of outcrops. Pillow lava samples were taken from pillow interiors; heavily veined, altered or amygdaloidal samples were avoided where possible.

In the laboratory, rock samples were reduced to pea-sized fragments in a chipmunk jaw crusher. After homogenization, approximately 200 grams of chips were pulverized in a tungsten-carbide swing mill.

Major-element oxides as well as Cu, Zn and Ni were determined on the rock powders in the Newfoundland Department of Mines and Energy, Mineral Development Division laboratory in St. John's. FeO was determined by titration with standard potassium dichromate and loss on ignition by heating to 1000°C. Other elements were determined by atomic absorption spectrometry. Methods used in these analyses are described in detail by Wagenbauer *et al.* (1983).

The remaining trace elements were determined by X-Ray fluorescence (XRF) at Memorial University of Newfoundland. For all except the rare-earth elements (REE), approximately 10 grams of sample powder were mixed with approximately 1.5 grams of bakelite resin, pressed to form a pellet and baked at 200°C for 15 minutes to set the binder. Pellets were then analyzed with a Phillips 1450 X-ray fluorescence spectrometer. A monitor saturated with trace elements was calibrated against standards and every element ratioed to it to correct for inter-run instrument drift.

Rare-earth elements were analyzed by X-ray fluorescence via a variant of the ion-exchange, thin-film method of Fryer (1977). Approximately 1.5 gm of sample was dissolved in HF and perchloric acid in teflon beakers. Samples were passed through ion exchange columns in 1.5N HNO₃ to remove barium and then through a second set in 2N HCl to isolate the rare-earth elements. The REE-bearing solution was enriched by evaporation, taken up on an ion exchange paper disk, dried and analyzed by XRF calibrated against international rock standards. All samples were analyzed in duplicate and replicate analyses confirm an accuracy of $\pm 10\%$ as quoted by Fryer (1977).

APPENDIX 2

PRECISION AND ACCURACY OF GEOCHEMICAL ANALYSES

Precision (P) for all elements is estimated from replicate analyses of blind duplicate pairs of samples. The precision is calculated as follows:

$$P = \frac{(V_1 - V_2) \times 100}{V_{1,2}}$$

V_1 = larger value of element pair

V_2 = smaller value of element pair

$V_{1,2}$ = mean of V_1 and V_2

The reported figure for P is the mean of all sample pairs analysed.

Accuracy of individual element analyses can be estimated from mean values for international standards analyzed by the same methods at the same time. Major element oxides are reported in weight percent, trace elements in ppm.

TABLE A-1: Detection limits for trace elements analysed by X-ray fluorescence

Element	Detection Limit (ppm)
Ba	20
Cr	10
V	10
Cu	5
Ni	5
Zn	5
Ga	5
Rb	2
Sr	2
Y	10
Zr	10
Nb	5

TABLE A-2: Precision (P) of major-element analyses

Element Oxide	P(mean)	St.Dev.	n
SiO ₂	0.62	0.49	12
Al ₂ O ₃	0.84	0.57	12
Fe ₂ O ₃	6.40	1.99	12
FeO	0.46	0.46	12
MgO	1.51	1.09	12
CaO	1.10	0.73	12
Na ₂ O	0.75	0.53	12
K ₂ O	7.10	10.85	12
TiO ₂	9.40	5.93	12
MnO	0.00	0.00	12
P ₂ O ₅	21.94	23.96	12
LOI	5.40	3.29	12

Note: St.Dev. = standard deviation; n = number of pairs

TABLE A-3: Precision (P) of trace-element analyses

Element	P(Mean)	St.Dev.	n
Rb	36.85	45.54	12
Sr	1.75	2.31	12
Y	13.81	10.60	12
Zr	8.49	7.21	12
Nb	ND	ND	12
Ba	34.76	21.65	12
V	7.42	7.02	12
Cr	0.92	0.93	12
Cu	4.99	4.93	12
Zn	0.84	1.01	12
Ni	3.50	1.86	12

Note: St.Dev. = standard deviation; n = number of pairs

TABLE A-4: Precision (P) of rare-earth-element analyses (10 pairs)

Element	P(Mean)	St.Dev.
La	5.71	3.64
Ce	5.03	1.86
Nd	5.37	3.51
Sm	6.71	5.00
Eu	17.19	10.88
Tb	30.35	30.81
Yb	13.06	18.87

Note: St.Dev. = standard deviation

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TABLE A-5: Accuracy of major-element analyses

AGV-1 Element Oxide	Abbey (1983)	<u>X</u>	<u>S</u>	<u>n</u>
SiO ₂	59.61	59.00	0.26	5
Al ₂ O ₃	17.19	16.89	0.24	5
Fe ₂ O ₃ *	6.83	6.69	0.09	5
MgO	1.52	1.46	0.02	5
CaO	4.94	4.76	0.09	5
Na ₂ O	4.32	4.16	0.05	5
K ₂ O	2.92	2.81	0.10	5
TiO ₂	1.06	1.01	0.02	5
MnO	0.10	0.09	0.01	5
P ₂ O ₅	0.51	0.54	0.06	5

BCR-1 Element Oxide	Abbey (1983)	<u>X</u>	<u>S</u>	<u>n</u>
SiO ₂	54.53	54.76	0.54	5
Al ₂ O ₃	13.72	13.53	0.17	5
Fe ₂ O ₃ *	13.33	13.46	0.06	5
MgO	3.48	3.48	0.03	5
CaO	6.97	6.86	0.13	5
Na ₂ O	3.30	3.25	0.04	5
K ₂ O	1.70	1.65	0.05	5
TiO ₂	2.26	2.03	0.18	5
MnO	0.18	0.19	0.00	5
P ₂ O ₅	0.36	0.38	0.01	5

G-2 Element Oxide	Abbey (1983)	<u>X</u>	<u>S</u>	<u>n</u>
SiO ₂	69.22	69.08	0.78	4
Al ₂ O ₃	15.40	15.26	0.16	4
Fe ₂ O ₃ *	2.65	2.60	0.01	4
MgO	0.75	0.75	0.01	4
CaO	1.96	1.85	0.03	4
Na ₂ O	4.06	4.08	0.04	4
K ₂ O	4.46	4.42	0.08	4
TiO ₂	0.48	0.53	0.05	4
MnO	0.03	0.03	0.01	4
P ₂ O ₅	0.12	0.13	0.01	4

MRG-1 Element Oxide	Abbey (1983)	<u>X</u>	<u>S</u>	<u>n</u>
SiO ₂	39.32	39.18	0.33	6
Al ₂ O ₃	8.50	8.56	0.12	6
Fe ₂ O ₃ *	17.74	18.16	0.35	6
MgO	13.49	13.46	0.12	6
CaO	14.77	14.52	0.43	6
Na ₂ O	0.71	0.69	0.02	6
K ₂ O	0.18	0.16	0.01	6
TiO ₂	3.69	3.68	0.28	6
MnO	0.17	0.18	0.01	6
P ₂ O ₅	0.06	0.08	0.04	6

Table A-5: Continued)

SY-2 Element Oxide	Abbey (1983)	<u>X</u>	<u>S</u>	<u>n</u>
SiO ₂	60.10	60.54	0.40	5
Al ₂ O ₃	12.12	12.14	0.31	5
Fe ₂ O ₃ *	6.26	6.24	0.04	5
MgO	2.70	2.63	0.10	5
CaO	7.98	7.88	0.13	5
Na ₂ O	4.34	4.29	0.04	5
K ₂ O	4.48	4.42	0.05	5
TiO ₂	0.14	0.15	0.02	5
MnO	0.32	0.32	0.01	5
P ₂ O ₅	0.51	0.44	0.02	5

Note: X = mean, S = standard deviation, n = number of samples; values in weight percent.

Table A-6: Accuracy of trace-element analyses by XRF

MRG-1 Element Oxide	Abbey (1983)	<u>X</u>	<u>S</u>	<u>n</u>
Rb	8	8.6	2.71	17
Sr	260	263.5	456	17
Y	16 ?	15.2	1.89	17
Zr	105	106.2	2.53	17
Nb	20 ?	25.2	2.54	16
Ba	50 ?	170.7	24.60	17
V	520	522.4	7.28	17
Cr	450	352.9	4.79	17

SY-2 Element Oxide	Abbey (1983)	<u>X</u>	<u>S</u>	<u>n</u>
Rb	220	224.4	3.81	23
Sr	275	262	3.71	23
Y	130	165.5	3.29	23
Zr	280	285.6	4.81	23
Nb	23 ?	35.3	2.84	23
Ba	460	434.2	13.41	23
V	52	44.9	3.44	23
Cr	12	5	0	23

Note: X = mean, S = standard deviation, n = number of samples; values in ppm.

TABLE A-7: Accuracy of trace elements analyzed by atomic absorption

<u>MRG-1</u>				
<u>Element Oxide</u>	<u>Abbey (1983)</u>	<u>X</u>	<u>S</u>	<u>n</u>
Cu	135	116	6	6
Ni	195	151	8	6
Zn	190	176	20.6	6
<u>SY-2</u>				
<u>Element Oxide</u>	<u>Abbey (1983)</u>	<u>X</u>	<u>S</u>	<u>n</u>
Ni	10	3.3	1	5
Zn	250	237	10.9	5

Note: X = mean, S = standard deviation, n = number of samples; values in ppm.

TABLE A-8: Accuracy of rare-earth-element analyses

<u>MRG-1</u>		
<u>Element Oxide</u>	<u>Abbey (1983)</u>	<u>stet.</u>
La	10 ?	11
Ce	25 ?	37.9
Nd	19 ?	26.1
Sm	5 ?	6.1
Eu	1.4 ?	1.4
Tb	-	0.6
Yb	1 ?	0.1

APPENDIX 3

WHOLE-ROCK GEOCHEMICAL DATA

TABLE A-9: Whole-rock geochemical data for the Pipestone Pond Complex

	Peridotite			Pyroxenite			
Sample No.	2140382	2140390	2140401	2140383	2140393	2140399	2140399
Zone	21	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	554580	573050	572620	569480	568570	566050	566050
Northing	5342090	5370300	5368420	5364190	5362890	5356130	5363620
SiO ₂	35.9	38.7	38.4	50.1	51.1	52.7	54.7
Al ₂ O ₃	0.15	0.53	0.38	2.33	3.22	0.86	2.03
Fe ₂ O ₃	3.87	3.02	3.48	1.17	1.51	0.96	0.38
FeO	3.06	4.18	4.04	4.09	5.13	5.50	9.36
MgO	45.98	42.53	41.70	20.16	18.41	21.49	18.49
CaO	0.14	0.58	0.87	18.16	17.76	14.83	11.62
Na ₂ O	0.05	0.05	0.06	0.07	0.09	0.08	0.11
K ₂ O	0.01	0.02	0.01	0.01	0.01	0.03	0.02
TiO ₂	0.03	0.12	0.07	0.08	0.10	0.06	0.12
MnO	0.10	0.11	0.13	0.15	0.18	0.19	0.11
P ₂ O ₅	0.01	0.01	0.01	0.01	0.01	0.01	0.01
LOI	12.75	12.06	11.96	3.31	1.40	3.17	2.60
Total	102.05	101.91	101.11	99.64	99.02	100.88	99.55
Cu	5	13	14	153	281	11	280
Zn	30	33	32	28	24	29	46
Ni	2340	2130	2180	254	194	167	264
Rb							
Sr							
V							
Zr							
Nb							
Ga							
Ba							
Y							
Cr	378	772	465	* 2498	1524	1628	726

Note: Values for oxides in weight percent; values for trace elements and rare-earth elements in ppm.

Mg # is atomic Mg/(Mg + Fe²⁺). Fe²⁺ calculated as 90% total Fe.

TABLE A-9. (continued)

Diabase								
Sample No.	2140333	2140347	2140367	2140369	2140371	2140372	2140376	2140387
Zone	21	21	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	565460	568420	567340	565710	565800	566140	566890	564930
Northing	5365240	5369300	5355490	5364290	5364200	5366000	5356560	5361820
SiO ₂	49.8	47.6	52	50	50.5	49.9	52.4	62.2
Al ₂ O ₃	13.65	16.65	14.25	13	12.8	11.45	14.3	14.8
Fe ₂ O ₃	2.49	1.37	2.3	1.54	1.85	1.91	2.69	3.93
FeO	8.28	5.71	10.23	9.54	8.84	8.76	10.93	3.56
MgO	10.80	9.95	5.59	11.14	10.95	12.68	3.94	1.95
CaO	8.48	13.59	5.18	9.23	10.62	10.75	5.95	9.10
Na ₂ O	2.48	1.09	4.58	1.53	1.66	0.42	2.04	1.65
K ₂ O	0.53	0.09	0.10	0.13	0.09	0.03	0.03	0.03
TiO ₂	0.45	0.14	0.98	0.38	0.3	0.22	0.65	0.67
MnO	0.20	0.18	0.29	0.23	0.23	0.22	0.26	0.12
P ₂ O ₅	0.02	0.01	0.07	0.04	0.02	0.02	0.03	0.08
LOI	2.81	4.06	2.92	2.62	2.08	3.19	5.88	2.57
Total	99.99	100.44	98.49	99.38	99.94	99.55	99.10	100.66
Cu	67	173	33	27	86	66	17	36
Zn	59	33	182	67	62	61	116	35
Ni	137	55	11	111	95	186	1	1
Rb	8	2	3	6	4	2	3	4
Sr	64	157	129	55	62	49	131	105
Y	11	2	22	9	5	6	13	3
Zr	13	6	27	11	6	8	11	3
Nb	1	2	1	3	1	1	4	1
Ga	4	9	14	6	35	3	1	8
Ba	58	832	76	39	314	42	52	45
V	333	147	500	310	342	290	409	213
Cr	603	122	35	623	547	882	5	5
La		0.3				0.1	0.03	
Ce		7.9				2.0	0.6	
Nd		0.7				0.2	0.06	
Sm		0.2				0.05	0.01	
Eu		0.8				0.3	0.4	
Gd		0.4				0.15	0.07	
Tb		0.5				0.22	0.1	
Er		0.3				0.1	0.03	
Yb		0.2				0.1	0.01	
Mg#	0.67	0.74	0.65	0.67	0.71	0.37	0.35	

TABLE A-9. (continued)

Basalt					
Sample No.	2140320	2140350	2140352	2140353	2140355
Zone	21	21	21	21	21
NTS	12A.1	12A.8	12A.8	12A.8	12A.8
Easting	565710	566900	567510	568210	568070
Northing	5355040	5353500	5352690	5352420	5351810
SiO ₂	49.8	45.3	47.8	49.2	46.8
Al ₂ O ₃	13.85	14.05	13.90	13.20	14.60
Fe ₂ O ₃	1.52	0.54	3.39	7.63	4.07
FeO	9.63	7.56	10.13	6.75	8.4
MgO	6.13	6.75	6.52	6.05	5.58
CaO	7.92	9.27	6.40	5.44	8.03
Na ₂ O	4.11	3.42	3.78	4.19	3.98
K ₂ O	0.46	0.11	1.09	1.58	1.57
TiO ₂	1.44	1.05	1.49	1.63	1.38
MnO	0.20	0.18	0.25	0.18	0.21
P ₂ O ₅	0.12	0.09	0.14	0.14	0.14
LOI	3.77	10.07	3.65	2.61	4.18
Total	98.95	98.39	98.54	98.6	98.94
Cu	53	65	36	55	31
Zn	83	61	106	100	92
Ni	53	86	47	42	52
Rb	7	7	34	24	28
Sr	96	127	108	48	77
Y	38	31	41	46	42
Zr	81	59	87	88	73
Nb	7	2	3	5	5
Ga	8	12	10	13	9
Ba	153	88	198	337	171
V	381	309	477	453	357
Cr	141	276	98	49	149
La		2.0	3.2	3.1	0.4
Ce		6.6	10.9	9.9	9.6
Nd		0.7	1.0	0.9	1.0
Sm		0.2	0.4	0.4	0.3
Eu		0.6	1.1	1	0.9
Gd		0.4	0.5	0.55	0.5
Tb		0.4	0.7	0.7	0.6
Er		0.2	0.4	0.4	0.3
Yb		0.2	0.3	0.3	0.3
Mg#	0.52	0.62	0.49	0.47	0.48

TABLE A-9. (continued)

Plagiogranite							
Sample No.	2140323	2140329	2140335	2140345	2140375	2140397	2140398
Zone	21	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	567900	565290	565510	562880	566890	564450	564450
Northing	5356490	5359190	5364290	5352300	5356560	5361700	5361700
SiO ₂	73	71.2	73.4	64.2	70.8	72.4	71.7
Al ₂ O ₃	12.9	12.35	12.6	17.6	12.75	13.5	13.15
Fe ₂ O ₃	0.50	0.40	1.18	0.26	0.50	0.92	0.70
FeO	3.65	4.81	2.49	4.95	4.16	2.78	3.01
MgO	1.20	2.01	0.86	1.77	1.29	0.74	0.78
CaO	1.51	1.09	2.58	1.35	1.88	3.69	3.34
Na ₂ O	5.04	2.81	4.32	6.85	4.81	3.60	3.67
K ₂ O	0.17	0.62	0.24	0.61	0.14	0.28	0.28
TiO ₂	0.30	0.35	0.19	0.65	0.31	0.26	0.27
MnO	0.07	0.07	0.05	0.12	0.13	0.09	0.09
P ₂ O ₅	0.02	0.06	0.02	0.16	0.05	0.04	0.04
LOI	1.62	2.99	1.85	1.93	2.41	1.58	1.62
Total	99.98	98.76	99.78	100.44	99.23	99.88	98.65
Cu	19	16	26	6	30	8	9
Zn	26	75	24	24	61	47	49
Ni	1	1	1	1	1	1	1
Rb	3	15	2	8	2	4	4
Sr	70	65	96	213	75	103	99
Y	30	16	22	69	34	26	25
Zr	42	17	12	172	26	25	23
Nb	1	1	1	4	1	1	1
Ga	7	9	19	16	5	9	8
Ba	65	60	178	284	46	88	82
V	36	57	28	14	56	22	24
Cr	5	5	5	5	5	5	5

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TABLE A-9. (continued)

Gabbro									
Sample No.	2140322	2140327	2140332	2140346	2140366	2140370	2140373	2140385	2140388
Zone	21	21	21	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	565460	568420	567340	565710	565800	566140	566890	564930	565910
Northing	5365240	5369300	5355490	5364290	5364200	5366000	5356560	5361820	5364190
SiO ₂	48.5	56.2	39.9	59.6	45.1	47.5	48.7	52.6	52.9
Al ₂ O ₃	16.9	14.15	16.55	14.90	16.6	17.05	18.20	14.80	13.80
Fe ₂ O ₃	1.17	1.67	7.46	1.54	2.69	1.84	0.80	2.56	1.79
FeO	4.66	10.70	10.42	6.58	7.63	5.4	4.83	8.36	9.22
MgO	9.97	4.31	7.83	3.92	8.63	10.07	10.31	5.62	8.16
CaO	10.74	3.36	12.30	5.16	14.32	14.25	10.84	9.70	8.98
Na ₂ O	2.66	4.50	0.45	3.22	0.89	0.44	1.72	2.06	2.02
K ₂ O	0.27	0.12	0.3	0.12	0.02	0.03	0.27	0.05	0.08
TiO ₂	0.13	0.67	0.55	0.31	0.20	0.13	0.04	0.36	0.20
MnO	0.13	0.19	0.20	0.18	0.24	0.18	0.15	0.25	0.20
P ₂ O ₅	0.10	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
LOI	4.50	2.94	2.3	3.11	3.06	3.23	4.07	2.63	2.38
Total	99.64	98.82	98.27	98.55	99.39	100.13	99.94	99.17	99.74
Cu	53	33	156	50	57	38	21	35	315
Zn	31	90	66	83	61	35	28	85	63
Ni	34	5	15	13	67	29	56	15	36
Rb	9	8	5	3	7	4	7	6	8
Sr	161	58	116	71	73	108	104	47	55
Y	1	8	1	12	3	3	1	5	1
Zr	1	8	4	10	7	5	3	2	3
Nb	1	1	1	1	1	1	1	1	1
Ga	6	5	11	6	7	26	4	6	7
Ba	69	59	96	108	30	23	51	10	43
V	151	337	1382	210	230	177	107	327	334
Cr	135	10	5	39	134	64	237	56	222
Mg#	0.78	0.41	0.48	0.49	0.63	0.74	0.79	0.51	0.60

TABLE A-9. (continued)

Gabbro		
Sample No.	2140389	2140391
Zone	21	21
NTS	12A/1	12A/8
Easting	565980	567650
Northing	5364190	5356900
SiO ₂	49.6	47.1
Al ₂ O ₃	9.90	16.45
Fe ₂ O ₃	1.06	1.49
FeO	8.16	5.93
MgO	14.31	11.67
CaO	12.00	11.59
Na ₂ O	0.96	1.38
K ₂ O	0.03	0.04
TiO ₂	0.24	0.11
MnO	0.24	0.18
P ₂ O ₅	0.01	0.01
LOI	3.63	4.09
Total	100.14	100.04
Cu	79	19
Zn	49	44
Ni	80	45
Rb	6	4
Sr	21	77
Y	3	2
Zr	3	4
Nb	1	4
Ga	1	10
Ba	29	42
V	270	164
Cr	380	246
Mg#	0.76	0.76

PIPESTONE POND AREA

TABLE A-10: Whole-rock geochemical data for the Cold Spring Pond Formation

North Salmon Dam Basalt								
Sample No.	2140305	2140307	2140308	2140311	2140330	2140331	2140336	2140337
Zone	21	21	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	562000	562500	562710	559420	563610	563590	563310	563020
Northing	5353210	5354900	5355180	5348520	5364200	5364170	5362790	5361690
SiO ₂	41.8	49.1	48.7	47.5	49.2	46.1	47.4	49.7
Al ₂ O ₃	10.40	14.25	14.05	14.00	14.05	14.05	11.40	4.30
Fe ₂ O ₃	2.95	4.09	2.40	4.59	2.79	1.48	1.86	3.87
FeO	8.19	8.14	8.77	8.19	6.98	12.36	9.97	14.42
MgO	12.98	5.04	8.24	7.59	5.60	4.58	11.38	11.51
CaO	14.84	9.06	10.66	9.57	9.56	9.96	9.71	11.43
Na ₂ O	1.63	4.85	3.38	3.81	3.90	3.61	1.70	0.36
K ₂ O	0.40	0.17	0.15	0.43	0.54	0.36	1.19	0.19
TiO ₂	1.60	2.37	2.13	2.13	2.44	3.41	1.46	0.71
MnO	0.20	0.18	0.22	0.21	0.19	0.23	0.28	0.35
P ₂ O ₅	0.18	0.22	0.20	0.26	0.11	0.32	0.19	0.06
LOI	3.20	1.53	1.10	2.00	3.62	2.35	3.12	1.37
Total	98.37	99	100	100.38	98.98	98.82	99.66	98.27
Cu	8	123	17	60	70	126	5	211
Zn	155	82	78	112	76	113	135	99
Ni	396	93	202	115	158	26	288	1253
Rb	5	5	5	3	19	8	30	7
Sr	185	200	228	284	232	273	135	13
Y	21	33	26	31	31	32	21	7
Zr	103	150	127	137	160	183	83	42
Nb	15	19	13	17	19	25	11	7
Ga	10	13	13	14	19	10	10	1
Ba	102	90	68	71	193	128	243	47
V	189	278	287	310	106	424	237	117
Cr	618	101	397	270	237	36	550	1437
La	9.7		9.8				7.9	
Ce	25.0		26.6				21.5	
Nd	14.7		20.1				15.1	
Sm	3.7		4.6				3.6	
Eu	1.3		1.5				1.0	
Gd	4.0		5.0				4.1	
Tb	0.6		0.6				0.7	
Yb	1.1		1.3				0.9	
Mg#	0.70	0.46	0.60	0.55	0.54	0.40	0.66	0.56

Notes: Values for oxides in weight percent; values for trace elements and rare-earth elements in ppm.

Mg# is atomic Mg/(Mg + Fe²⁺). Fe²⁺ calculated as 90% total Fe. NA—Not analysed.

TABLE A-10. (continued)

North Salmon Dam Basalt

Sample No.	2140338	2140363	2140780	2140781	2140782	2140783	2140784
Zone	21	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	563450	561380	560950	550550	562360	561850	569680
Northing	5360210	5351000	5350500	5350120	5353700	5353080	5348450
SiO ₂	48.7	43.3	48.6	50.35	51.05	42.65	48.55
Al ₂ O ₃	13.15	10.75	13.87	14.04	12.71	9.72	13.22
Fe ₂ O ₃	2.56	5.99	3.71	2.29	2.01	1.64	3.12
FeO	9.85	5.56	5.26	9.2	11.54	9.19	8.98
MgO	8.02	8.57	2.72	5.52	5.69	12.05	7.40
CaO	8.21	19.46	15.21	9.44	9.87	16.15	9.65
Na ₂ O	3.66	0.48	3.40	3.94	3.35	1.51	3.86
K ₂ O	0.36	0.13	0.09	0.19	0.33	0.38	0.52
TiO ₂	2.16	1.84	3.19	2.39	1.87	17.40	2.16
MnO	0.22	0.19	0.15	0.23	0.27	0.19	0.19
P ₂ O ₅	0.21	0.16	0.33	0.24	0.16	0.27	0.25
LOI	1.95	3.44	3.31	0.82	0.40	3.59	1.58
Total	99.05	99.87	99.84	98.72	99.25	99.08	99.48
Cu	283	6	25	83	10	4	59
Zn	101	73	75	101	107	167	122
Ni	181	487	34	91	45	413	131
Rb	11	9	1	6	9	17	10
Sr	282	607	503	204	123	197	255
Y	26	27	34	37	60	22	29
Zr	131	112	225	139	103	96	127
Nb	16	14	39	22	9	14	19
Ga	13	13	20	12	11	11	11
Ba	62	56	42	77	61	83	97
V	310	120	120	346	505	195	310
Cr	324	594	53	264	98	603	302
La	9.8						
Ce	29.3						
Nd	18.6						
Sm	5.0						
Eu	1.3						
Gd	5.6						
Tb	0.8						
Yb	1.3						
Mg#	0.57	0.61	0.39	0.49	0.46	0.69	0.55

TABLE A-10. (continued)

Felsic Volcanic Rocks						
Sample No.	2140736	2140737	2140738	2140740	2140741	2140743
Zone	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	559220	559270	559500	559570	559570	560550
Northing	5343200	5543450	5344250	5344450	5344450	5347440
SiO ₂	72.8	69.7	72.6	70.7	70.8	70.8
Al ₂ O ₃	11.60	13.75	11.95	13.9	13.75	13.95
Fe ₂ O ₃	0.75	0.75	0.82	0.81	0.74	0.84
FeO	2.02	2.5	2.29	1.99	2.16	1.98
MgO	1.17	1.25	1.70	1.23	1.10	1.33
CaO	1.00	0.58	0.63	0.48	0.52	0.48
Na ₂ O	1.49	1.94	1.1	2.23	2.20	1.88
K ₂ O	4.75	5.19	5.78	4.55	5.15	4.58
TiO ₂	0.47	0.51	0.48	0.48	0.53	0.51
MnO	0.04	0.04	0.02	0.04	0.05	0.04
P ₂ O ₅	0.18	0.2	0.17	0.19	0.18	0.18
LOI	2.13	1.82	1.83	2.04	1.76	2.05
Total	98.4	98.23	99.37	98.64	98.94	98.52
Cu	NA	NA	NA	NA	NA	NA
Zn	NA	NA	NA	NA	NA	NA
Ni	NA	NA	NA	NA	NA	NA
Rb	123	164	138	132	149	128
Sr	44	54	43	53	69	55
Y	41	42	37	45	47	55
Zr	162	176	160	186	192	185
Nb	15	17	13	17	13	16
Ga	15	20	18	22	20	20
Ba	583	699	868	596	772	605
V	30	37	30	39	37	39
Cr	17	18	12	17	14	13

TABLE A-11: Whole-rock geochemical data for the North Steady Pond Formation (Baie d'Espoir Group)

Felsic Volcanic Rocks								
Sample No.	2140340	2140341	2140342	2140351	2140356	2140358	2140359	2140381
Zone	21	21	21	21	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8	12A/8
Easting	565600	566080	565650	555720	566820	567720	567790	562980
Northing	5351000	5349490	5350540	5349020	5349600	5349400	5347600	5351510
SiO ₂	68.6	65	65.8	71.8	62.2	69.7	67.7	85.8
Al ₂ O ₃	14.50	14.70	15.75	13.40	13.10	15.05	14.60	8.90
Fe ₂ O ₃	3.65	0.99	0.28	0.13	0.52	1.05	3.04	0.07
FeO	0.00	4.47	3.83	2.97	5.71	1.25	1.08	0.14
MgO	0.48	2.16	0.78	1.01	2.47	1.40	0.74	0.17
CaO	0.44	4.31	0.65	0.36	3.8	0.56	1.01	0.16
Na ₂ O	2.01	3.09	1.83	2.19	3.06	0.48	3.08	4.66
K ₂ O	8.09	1.01	8.71	5.47	1.88	6.07	5.17	0.12
TiO ₂	0.33	0.87	0.34	0.29	0.81	0.28	0.43	0.15
MnO	0.05	0.10	0.04	0.03	0.13	0.04	0.03	0.01
P ₂ O ₅	0.05	0.14	0.06	0.06	0.14	0.17	0.07	0.01
LOI	1.11	2.32	1.52	2.79	4.83	3.06	1.53	0.66
Total	99.31	99.16	99.59	99.5	98.66	99.11	98.43	100.85
Cu	12	7	5	16	9	10	4	3
Zn	60	67	66	55	73	25	48	4
Ni	1	4	2	1	5	1	1	1
Rb	188	61	246	171	76	214	139	3
Sr	48	125	94	58	165	65	137	45
Y	69	50	77	58	46	54	61	70
Zr	302	229	311	265	213	139	323	87
Nb	15	14	16	13	12	13	15	1
Ga	14	16	24	13	15	17	14	1
Ba	817	253	860	796	554	1588	844	77
V	21	103	15	20	106	15	12	10
Cr	5	5	5	5	5	5	5	5

Notes: Values for oxides in weight percent; values for trace elements in ppm.

TABLE A-12: Whole-rock geochemical data for the Sliver Basalts

'Sliver Basalts'				
Sample No.	2140328	2140360	2140379	2140785
Zone	21	21	21	21
NTS	12A/1	12A/8	12A/8	12A/8
Easting	565490	565120	554120	554160
Northing	5358080	5342120	5340740	5340660
SiO ₂	48.8	50	49.9	48.7
Al ₂ O ₃	16.20	14.65	13.70	13.52
Fe ₂ O ₃	7.87	0.52	2.52	2.36
FeO	6.83	8.08	8.53	7.98
MgO	3.30	3.73	6.79	6.59
CaO	5.08	7.42	10.52	12.15
Na ₂ O	5.48	4.21	3.14	3.35
K ₂ O	1.02	0.81	0.13	0.08
TiO ₂	1.43	0.91	1.16	1.10
MnO	0.19	0.17	0.20	9.19
P ₂ O ₅	0.11	0.05	0.11	0.08
LOI	2.71	8.43	2.99	4.11
Total	99.02	98.98	99.69	100.21
104	104	100	70	30
Zn	77	84	80	77
Ni	51	12	49	48
Rb	22	17	7	7
Sr	71	195	87	81
Y	36	22	36	28
Zr	73	25	64	41
Nb	5	3	4	6
Ga	7	13	12	9
Ba	93	360	167	22
V	474	454	350	314
Cr	115	61	118	132
Mg#	0.32	0.46	0.55	0.56

Notes: Values for oxides in weight percent; values for trace elements in ppm.

Mg# is atomic Mg/(Mg + Fe²⁺). Fe²⁺ calculated as 90% total Fe.

**MAPS
NOT
FILMED**

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